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Crustal structure of the NE Rockall Trough from wide-angle seismic data modeling

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[1] Two wide-angle seismic lines located in the northern Rockall Trough were acquired in May 2000. One line (line E) crosses the trough from the continental shelf off Lewis to normal oceanic crust west of Lousy Bank in NW-SE direction. The other line (line D) intersects with line E, crosses the Wyville-Thomson Ridge in a SW-NE direction and ends in the Faeroe-Shetland Basin. Sonobuoy data and expanding spread profiles acquired in the same area have been remodeled. Analysis of the seismic data using travel times and amplitudes reveals an up to 5 km thick sedimentary basin including an up to 1.5 km thick basaltic layer which is present in most of the trough. Further conclusions of this study are that the Rockall Trough is underlain by highly stretched continental crust of ~13 km thickness. The crust thickens to ~24 km beneath Lousy Bank, which is interpreted to be of continental nature. Beneath the Hebrides continental shelf a three-layer continental crust of 26 km is modeled. An up to 12 km thick high-velocity layer is observed underneath the ocean-continent boundary and is interpreted as magmatic underplating resulting from excess volcanism during rifting. No evidence for an underplate layer could be distinguished beneath the trough area. Modeling of the structure of the Wyville-Thomson Ridge revealed no existing igneous core of the ridge confirming existing theories, that it is a compressional structure.

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1. Introduction

[2] The Rockall Trough is a 250 km wide and up to 3 km deep bathymetric basin separating the Irish and UK continental shelves from the Rockall Bank (Figure 1). It is one of a series of rift basins that formed prior to the opening of the present-day North Atlantic Ocean. To the north the Rockall Trough is separated from the Faeroe-Shetland Trough by the NE Rockall Basin and the Wyville-Thomson Ridge. Early Tertiary continental breakup in the northern North Atlantic was accompanied by massive extrusive volcanism due to the presence of the Iceland plume, as evidenced by a thick layer of flood basalts in the northern Rockall Trough [Barton and White, 1997]. In the last 10 years this region has been the focus of exploration for hydrocarbons which has required knowledge of the nature and thickness of the

crust beneath the trough and its rifting history. The new data presented here provide constraints on the nature and thickness of the crust and further knowledge of the development of aulacogens or failed attempts at rifting to form an ocean basins.

[3] The age of the Rockall Trough has been subject to discussion. Ages derived from review of adjacent basins and plate reconstruction studies cover a wide range from the Torridonian [Roberts *et al.*, 1988], pre-Permian [Smythe, 1989], Late Carboniferous [Shannon *et al.*, 1995] to Cretaceous [Musgrove and Mitchener, 1996; England and Hobbs, 1997]. Evidence for pre-Cretaceous rifting in the Rockall Trough has been found from two-dimensional (2-D) structural and stratigraphic forward modeling [Nadin *et al.*, 1999]. The rift system might have continued into the Faeroe-Shetland Basin until the formation of the Wyville-Thomson Ridge through inversion in the mid-Tertiary [Boldreel and Andersen, 1993; Tate *et al.*, 1999].

[4] The continental nature of the Rockall Plateau was first deduced from plate reconstructions [Bullard *et al.*, 1965]. This hypothesis has been subsequently verified by wide-angle seismic profiling [Scrutton, 1970; Bunch, 1979]. The crustal structure of deeper parts of the Rockall Trough has been derived mainly from wide-angle seismic data. In the absence of high-quality deep crustal data, early work

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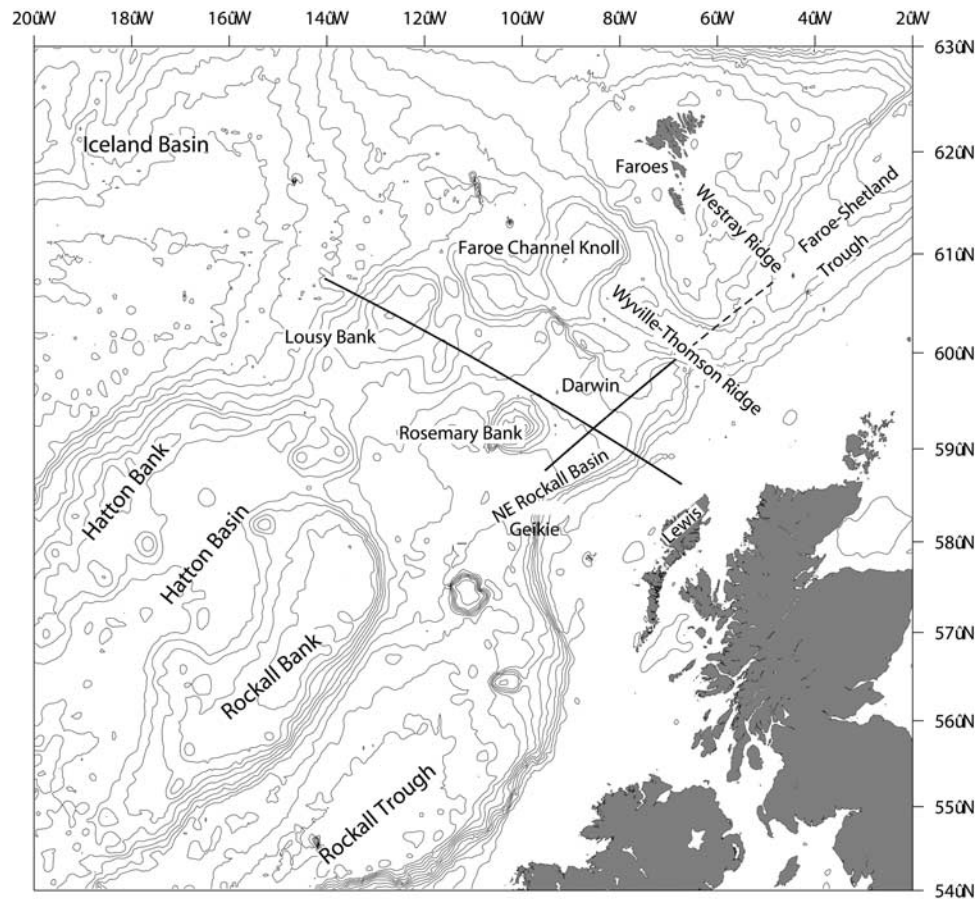


Figure 1. Continental margin west of Britain. Bold lines show wide-angle profiles, and major physiographic features are shown. Contours (250 m contour interval) are predicted seafloor bathymetry from satellite gravity [Smith and Sandwell, 1997].

assumed oceanic crust in the middle of the trough [e.g., Talwani and Eldholm, 1977]. Analysis of two wide-angle seismic profiles crossing the Rockall Trough at 57°N and 58.5°N suggests that the deep basin is underlain by continental crust [Roberts *et al.*, 1988]. The two-layered continental crust is asymmetrically thinned in the center of the trough, and the midcrustal discontinuity is at its minimum depth closer to the thicker crust beneath the east margin and above the shallowest elevation of the Moho [Roberts *et al.*, 1988]. Data from two-ship seismic experiments in the center of the Rockall Trough show a roughly 6 km thick crust [Joppen and White, 1990]. Joppen and White conclude, also based on the interpretation of the two wide-angle profiles further to the north [Roberts *et al.*, 1988], that the crust in the area of the two-ship experiment is also of continental affinity. They do not exclude that part of the crust might be oceanic or might consist of intrusions from decompression melting [Joppen and White, 1990].

[5] Using a seafloor cable recording explosive shots, the Pull-Up Multichannel Array (PUMA) experiment confirmed that the three layered continental crust west of Lewis is about 26 km thick with either a sharp Moho discontinuity or a zone of high- and low-velocity laminations [Powell and Sinha, 1987]. Between 1987 and 1989 the United Kingdom Department of Trade and Industry acquired combined reflection, wide-angle and expanding spread seismic data

from the UK mainland across the Hebrides Platform, Rockall Trough, and Rockall Plateau. Analysis of this data set revealed a velocity structure extending over the entire area of investigation which is consistent with continental crustal rocks, thinning from about 26 km underneath the UK mainland and Rockall Bank to 13.7 km in the Rockall Trough [Neish, 1993]. The crust is two-layered with an upper crustal layer of constant thickness across the Rockall Trough and pinching out at the ocean-continent boundary and a lower crust thickening underneath the shelf and underneath Rockall Bank [Neish, 1993].

[6] Modeling of wide-angle and reflection seismic data in conjunction with expanding spread profiles revealed two series of seaward dipping reflectors, consisting of submarine and subaerial lava flows and an up to 12 km thick high-velocity layer underneath the Hatton Bank continental margin, interpreted as igneous underplating [Morgan and Barton, 1990]. The Rockall and Porcupine Irish deep seismic project carried out several deep reflection seismic and wide-angle surveys in the Hatton Basin and continental margin region. The main results of the modeling were the confirmation of the continental affinity of the crust in the Hatton Basin and the existence of a zone of thick lower crust below the Hatton margin east of anomaly 24, interpreted as underplated material [Vogt *et al.*, 1998]. Both the underplating and the seaward dipping reflectors can be

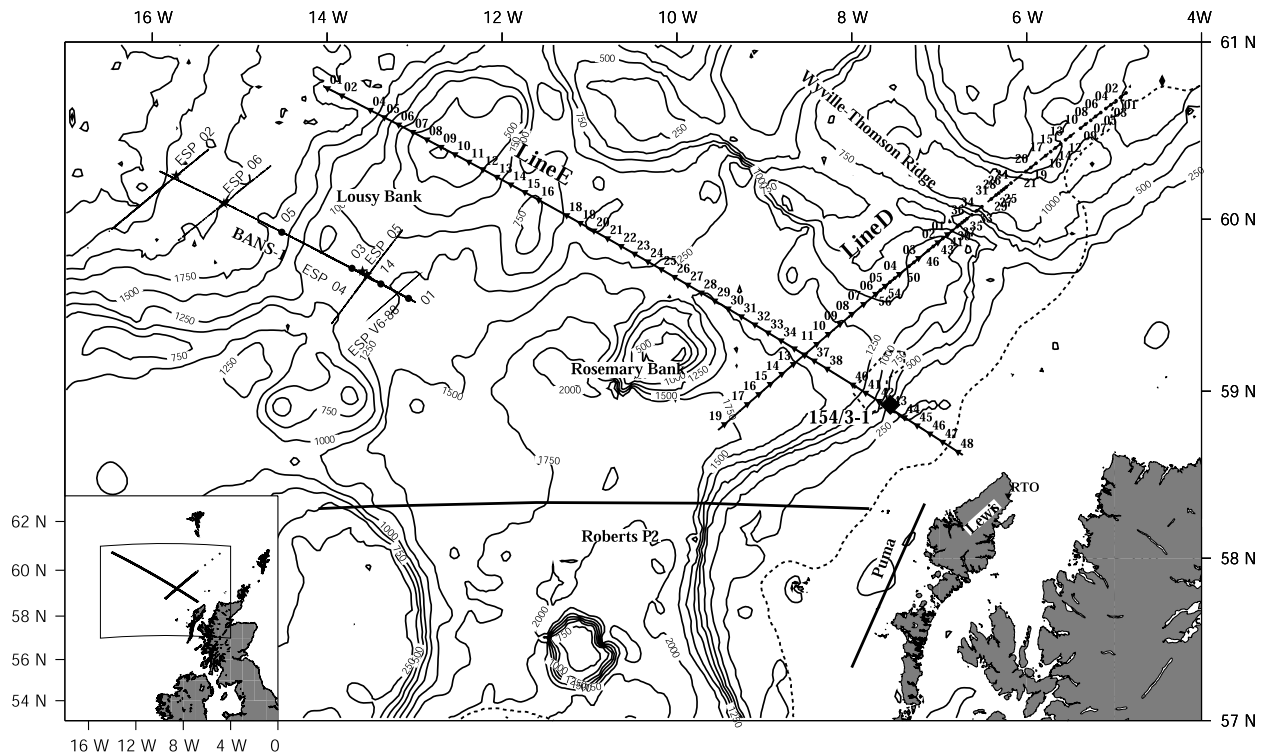


Figure 2. Location map of lines E and D and the BP/Shell data set. Inverted triangles show location of OBS recovered successfully on lines E and D. Circles show successful recovery locations of OBS from the BP/Shell line used to extend line D. Black dashed line marks the eastward extend of the basalt and the hole in the basalt [after *Stoker et al.*, 1993]. Black triangle shows British Geological Survey (BGS) land station location. Circles mark sonobuoy position on the BANS 1 profiles, and stars mark the midpoints of the ESP used for the modeling. Position of the BP drill hole 154/3-1 is marked by a diamond. Contours are predicted seafloor bathymetry from satellite gravity [Smith and Sandwell, 1997] in a 250 m interval.

explained by the volcanism caused by decompression melting of the passively upwelling asthenosphere underneath the thinning lithosphere during rifting. The main difference between volcanic and nonvolcanic margins is the potential temperature of the asthenosphere, which is up to 150° higher underneath volcanic margins [White *et al.*, 1987a]. Seismic methods have only detected underplating directly at the continent-ocean boundary, while subsidence analysis suggests regional underplating of 4–5 km thickness in the northern Rockall region [Brodie and White, 1994; Clift and Turner, 1998]. A layer consisting of velocities similar to igneous underplating underneath the trough has been interpreted as a zone of extensive serpentinization [O'Reilly *et al.*, 1996]. Subsequent remodeling of the data has not confirmed this interpretation [Pearse, 2002].

[7] It has been inferred that the Wyville-Thomson Ridge, Ymir Ridge, and Bill Bailey's Bank are compressional ramp anticlines located above fault planes [Boldreel and Andersen, 1994]. This has been based on the observation that basaltic lavas within these structures are folded and that sedimentary sequences of the same age display onlap on one side of the structure and are uplifted and truncated on the other side. Ridge push from the initiation of seafloor spreading in the North Atlantic has been proposed as cause for the NE-SW compressional stress necessary to form these

structures [Boldreel and Andersen, 1994]. Recent investigations using reflection seismic data acquired in the region of the NE Rockall Basin show that the Wyville-Thomson Ridge has no igneous core and only a thin basalt cover overlying 3–4 km thick sedimentary layers [Tate *et al.*, 1999]. On the basis of these new data, the Wyville-Thomson Ridge together with the Ymir and Munkegrunnur ridges can be considered as positive inversion ramp anticlines above an inferred mid-lower crustal detachment, generated during N-S intraplate compression [Tate *et al.*, 1999]. The origin of the deformation is proposed to be caused by compressional stress in the Pyrenean-Alpine foreland.

2. Seismic Data

[8] In May 2000 two wide-angle seismic lines were acquired by the Atlantic Margins Project (AMP) in the NE Rockall Trough (see Figure 2) using the R/V *Akademik Boris Petrov* and 50 ocean bottom seismographs (OBS) from Geopro GmbH. The main aims of the experiment were the study of the deep structure of the Rockall Trough and the ocean continent transition zone as well as detection and quantification of magmatic underplate in the study area.

[9] The main difficulty in imaging the deep structure of the crust in the Rockall Trough is the cover of basaltic lava flows, present across vast regions of the trough. Since basalt

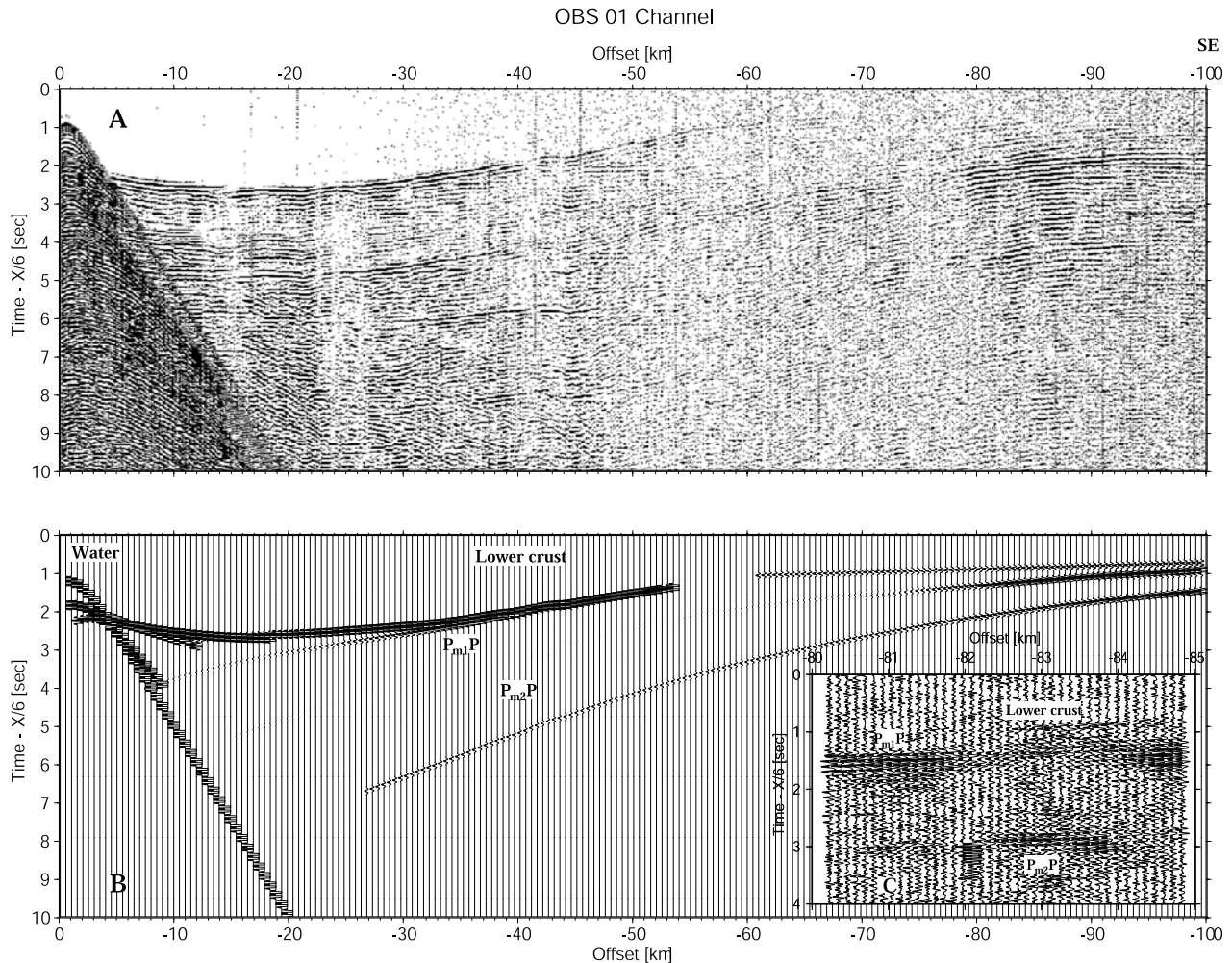


Figure 3. (a) Band-pass-filtered (3–5 Hz, 24–36 Hz) data from the hydrophone channel of OBS 01 (hydrophone) on line E. The data are gain-adjusted according to offset and reduced to a velocity of 6 km/s. (b) Synthetic seismograms calculated from the final velocity model for the same OBS using asymptotic ray theory code [Zelt and Smith, 1992]. The synthetic seismograms are spaced at a 500 m interval, and the same offset-dependent gain has been applied as to the seismic data. The wavelet consists of a 29-point low-passed Ricker wavelet. P wave quality factors were chosen to be 100–200 for sedimentary layers, 400 for the basaltic layers and upper crust, 700 for the lower crust, and 1000 for the upper mantle.

is not deposited homogeneously, scattering, reflection and attenuation of seismic energy are considerable. Furthermore, a basaltic layer often represents a high-velocity layer, where the point of critical reflection is quickly reached preventing energy from penetrating the basalt layer. Sedimentary layers underneath the basalt then represent low-velocity layers from which no energy is returned. The use of low frequencies and long offset data may help to overcome some of the difficulties. Therefore an experiment with ocean bottom seismometers was designed to record arrivals from a seismic source rich in low frequencies ensuring good penetration [Ziolkowski *et al.*, 2003].

[10] A total of 66 successful deployments were undertaken on the two profiles. The instrument spacing on both profiles was on average 10 km. All instruments recorded on one vertical, two horizontal geophones and a hydrophone channel. The tuned air gun array, with the main frequencies centered around 8 Hz, consisted of 10 air guns with a total volume of 3040 cubic inch (49.8 l), towed at a depth of

12 m. The shot interval was 60 s at an average speed of 4 knots, which translates to a trace spacing of about 120 m. The sample rate was 8 ms for all OBS. A total of 5576 shots (line E 3881, line D: 1695) were fired by the air gun array.

[11] Line E crosses the Rockall Trough from the continental shelf off Lewis extending over Lousy Bank and onto oceanic crust as defined by linear magnetic anomalies. It has a length of 480 km and 48 ocean bottom seismographs (OBS) from Geopro GmbH were used of which 45 yielded usable data. All shots on this profile were additionally recorded by British Geological Survey land stations, which allowed the profile to be extended to a length of 520 km using the data from land station RTO which lies on the onshore projection of the profile.

[12] Line D crosses line E and ties with an existing BP/Shell data set which has been remodeled by the AMP. It was acquired using 20 OBS of which 19 yielded usable data and one was lost during the experiment. The line length is

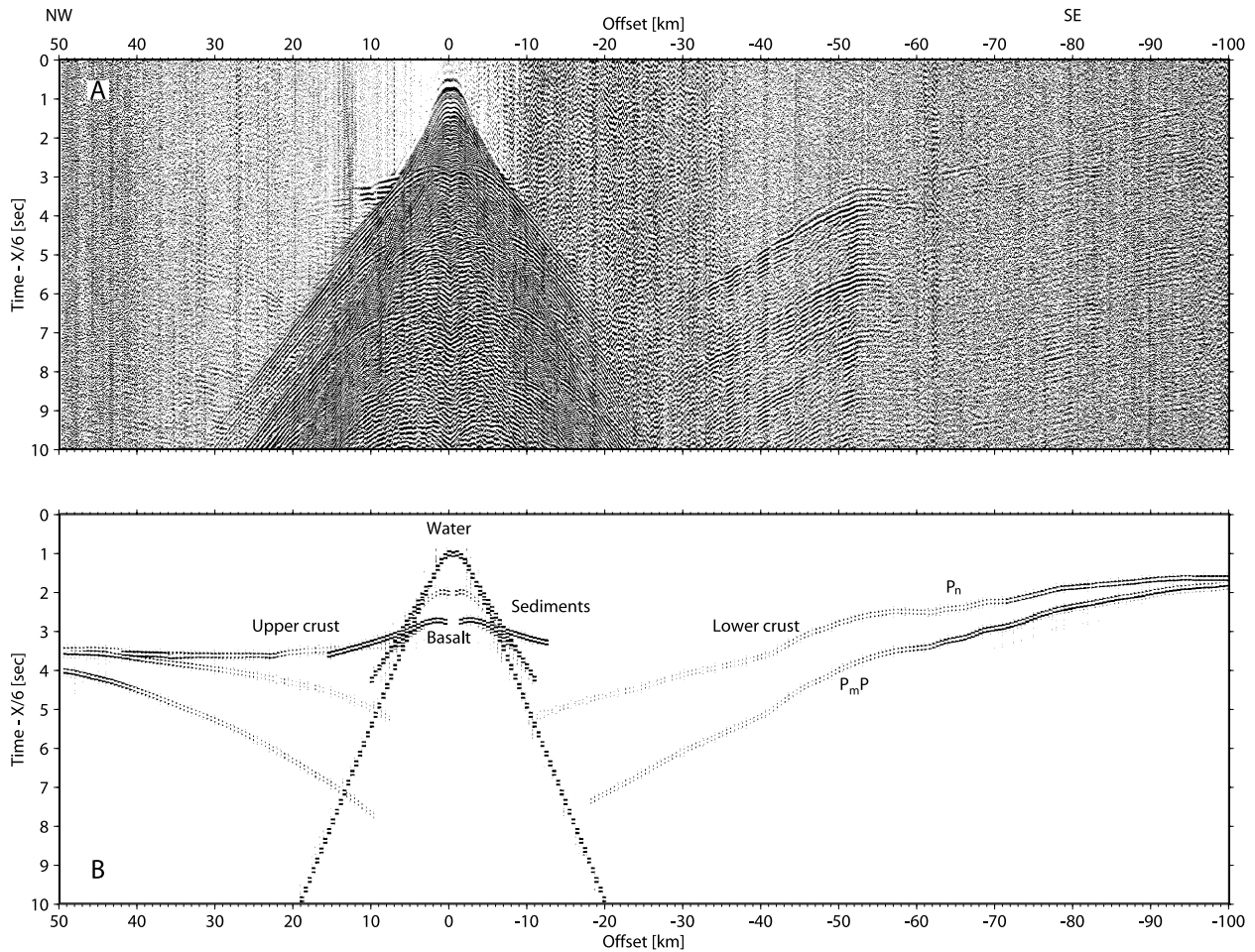


Figure 4. (a) Data from the hydrophone channel of OBS 38 for line E with same gain, filter, and scaling applied as in Figure 3 and (b) the corresponding synthetic seismograms.

180 km. The existing BP/Shell profile consists of 992 shots recorded by a total of 57 OBS, of which 45 provided useful data. The air gun array for the BP/Shell profile consisted of two air guns of 60 l each (a total of 7322 cubic inches). The same shot interval and sample rate were used as on line D

and E. The length of the combined D and BP/Shell profiles is 350 km.

[13] Additionally, the scope of the AMP experiment was enhanced by access to a previously unpublished data set of combined expanding spread profile (ESP), sonobuoy and

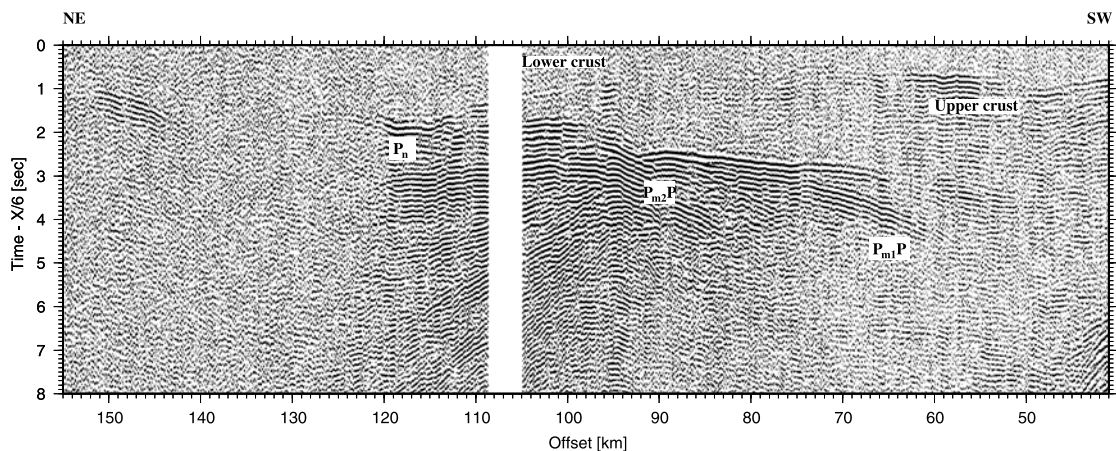


Figure 5. Data from vertical geophone channel of land station RTO. The data are gain-adjusted according to offset and reduced to a velocity of 8 km/s.

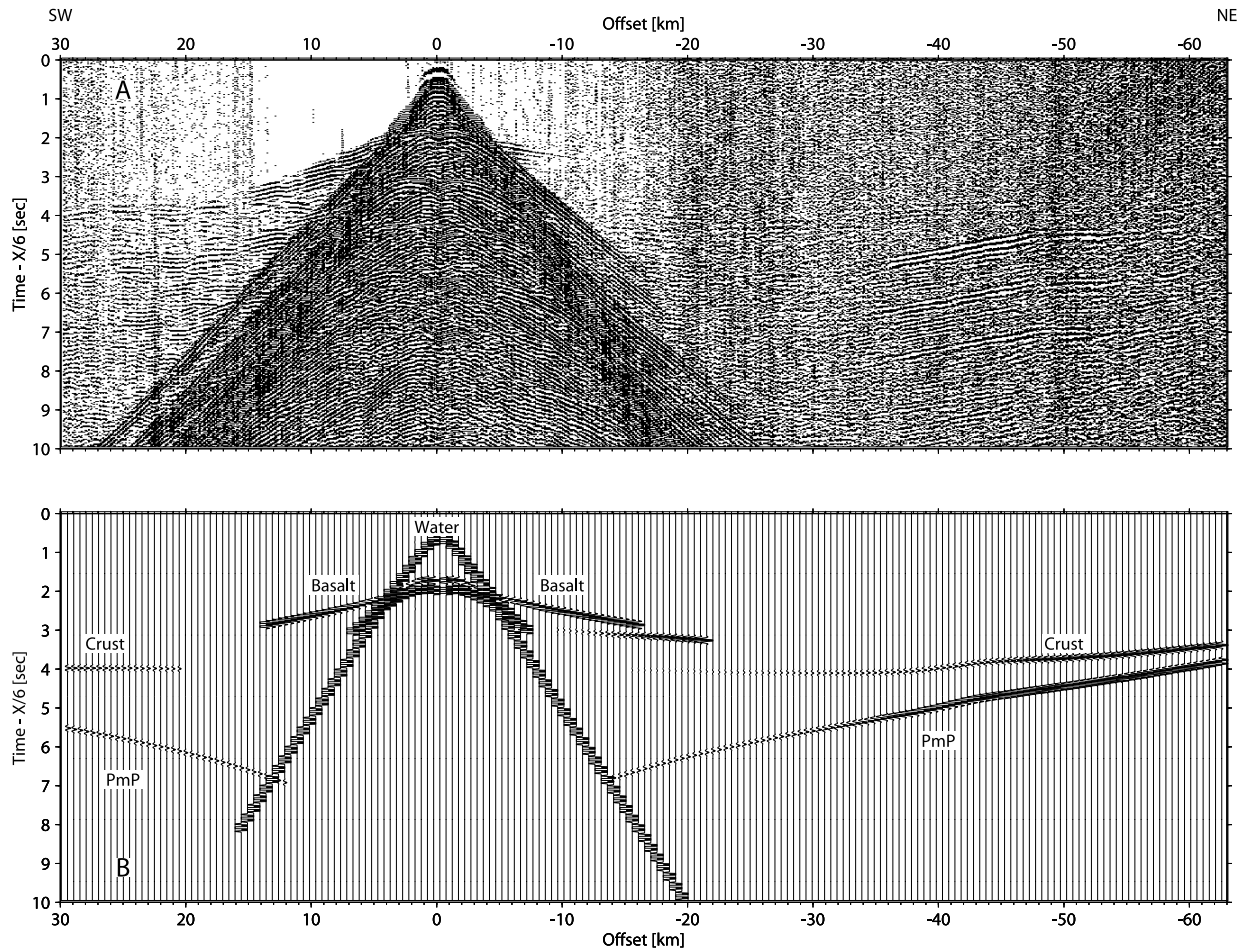


Figure 6. (a) Data from the hydrophone channel of OBS 06 for line D with same gain, filter, and scaling applied as in Figure 3 and (b) the corresponding synthetic seismograms.

reflection seismic data acquired by the U.K. Department of Energy (DTi) in the Rockall Trough and Hatton Trough (Figure 2) which were reprocessed. The BANS 1 reflection profile crosses the continental margin of the Rockall Trough south of AMP profile E at 59°N onto oceanic crust. Four ESP profiles oriented at 90° to the profile were used to constrain the deep velocity structure of the margin. A 7280 in³ (120 L) linear air gun array and a 4 km streamer with 300 channels were used during this survey. Four sonobuoys provided sufficiently long offsets (>25 km) of useful data. The seismic traces were redigitized from DTi paper sections to preserve amplitude information and included in the modeling.

3. Velocity Modeling

[14] Preprocessing of the data acquired for the Atlantic Margins Project undertaken by Geopro GmbH included calculation of the clock drift corrections to adjust the clock in each instrument to the GPS base time. Instrument locations were corrected for drift from the deployment position during their descent to the seafloor by minimizing the misfit between the predicted and recorded first water arrivals.

[15] Data quality is very good on the western part of line E overlying oceanic crust (Figure 3). Useful arrivals could be picked out to offsets of 120 km, including arrivals reflected from the top of a high velocity lower crustal layer and from the Moho. Because of the shallow water depth the data quality decreases over Lousy Bank, where multiples sometimes overlie secondary arrivals and strong reverberations can be observed in the vertical and horizontal channels of the OBS. In the Rockall Trough the data quality is variable, generally improving toward the eastern side of the trough where the basalt layer is thinner. The instruments located nearby and over a hole in the basalt at 380 km model distance yield high-quality data. Reflections from the Moho (*PmP* arrivals) can be traced to offsets of 80–100 km (Figure 4). Land station RTO (Figure 5) which has been included into the model recorded shots up to 150 km offset. Again due to very shallow water depth, the quality of the data on the continental shelf decreases toward the eastern end of the model.

[16] The data quality along line D is generally good (Figure 6) and improves toward the NE due to the thinning of the basaltic layer. Narrow bands of noise, possibly resulting from tidal currents are found on some sections. The quality of the BP/Shell data set is good throughout the

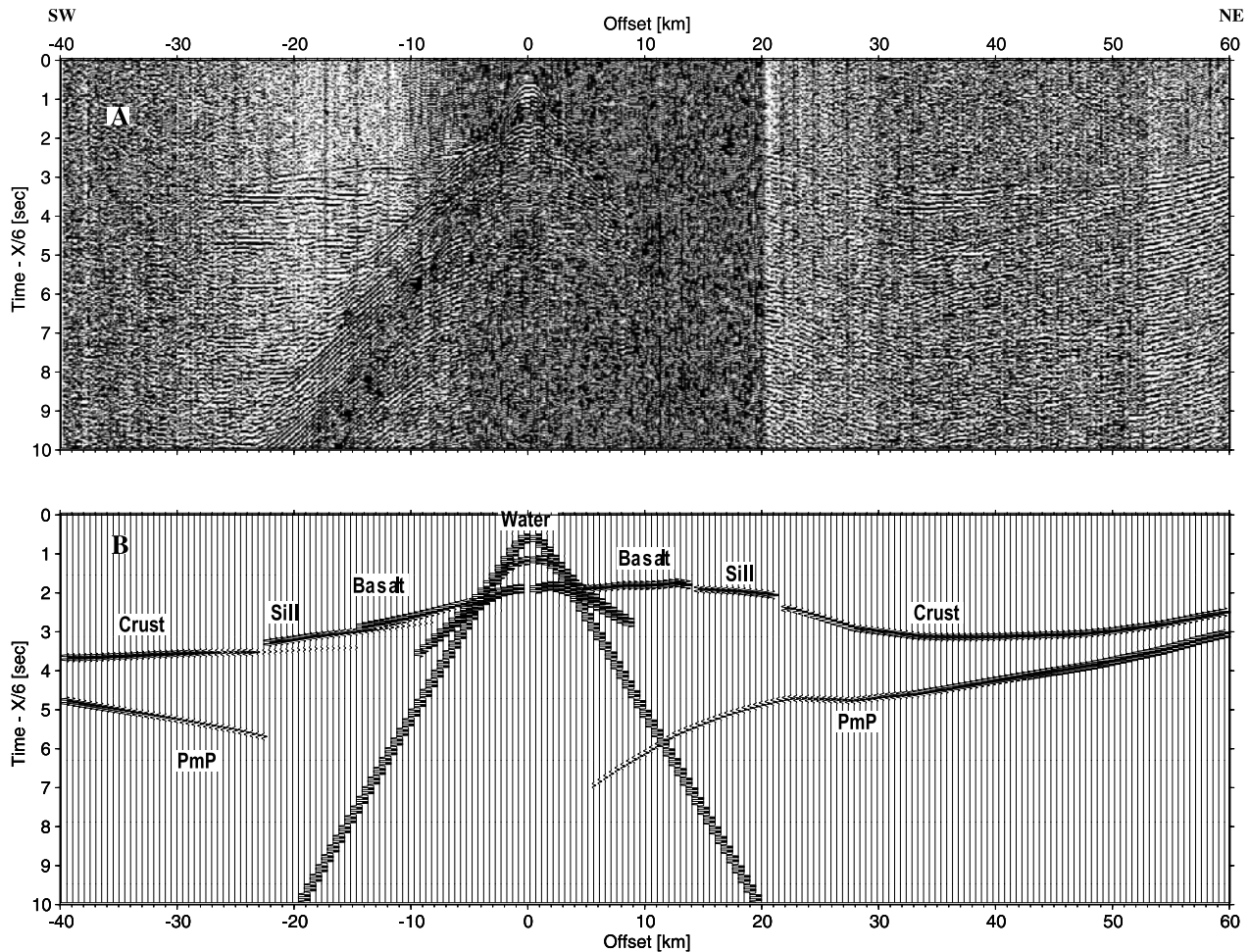


Figure 7. (a) Data from the hydrophone channel of OBS 42 from the BP/Shell data set with same gain, filter, and scaling applied as in Figure 3 and (b) the corresponding synthetic seismograms.

line. The close spacing of the instruments allows a good correlation of phases from one OBS to the next. Moho reflections or turning rays from the upper mantle were observed on 23 instruments along the line (Figure 7).

[17] The data were modeled using the inversion and ray tracing algorithm of [Zelt and Smith, 1992]. Modeling was performed using a layer-stripping approach, proceeding from the top of the structure toward the bottom. We used

a two-dimensional iterative damped least squares inversion of travel times [Zelt and Smith, 1992]. Upper layers, where not directly constrained by arrivals from within the layer, were adjusted to improve the fit of lower layers. For the model parameterization we used the minimum parameter/minimum structure approach to avoid inclusion of velocity or structural features into the model unconstrained by the data [Zelt, 1999]. Lateral velocity changes are included into

Table 1. Travel Time Residuals and Chi-Square Error for All Phases and the Complete Model of Line E

Phase	Number of Picks	RMS Travel Time Residual	Chi-Square
Water	1819	0.041	1.035
Sediments 1 reflection	1632	0.106	3.147
Sediments 2 reflection	826	0.078	1.687
Sediments 3	335	0.167	4.378
Basalt	1134	0.074	1.538
Upper crust	5731	0.095	5.599
Lower crust	1373	0.081	1.828
Midcrustal reflection (oc.)	305	0.206	4.263
PmP/top underplate	2219	0.199	3.962
Base underplate	691	0.128	2.562
Pn	149	0.264	10.995
All phases	16214	0.118	3.681

Table 2. Travel Time Residuals and Chi-Square Error for All Phases and the Complete Model of Line D

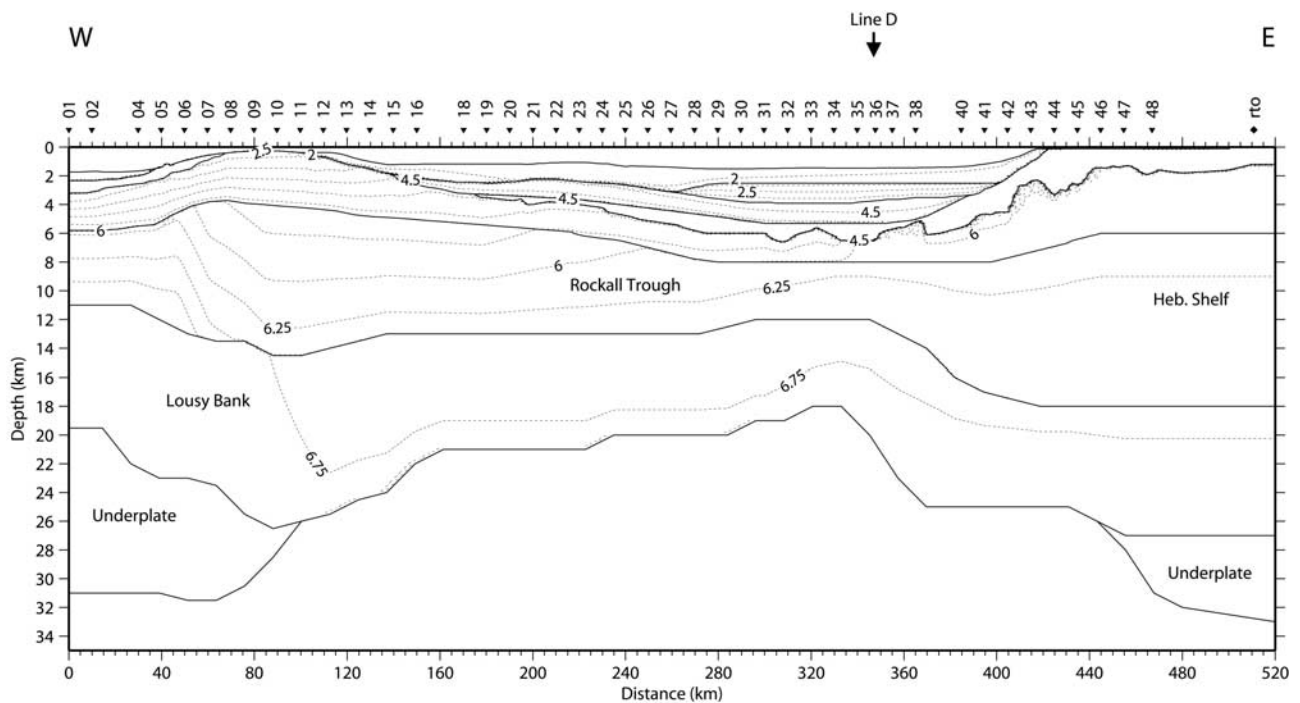
Phase	Number of Picks	RMS Travel Time Residual	Chi-Square
Water	2292	0.037	1.045
Sediments 1	1016	0.059	0.728
Sediments 2	403	0.050	0.709
Sediments 3	1751	0.110	2.441
Basalt	2520	0.073	1.630
Upper crust	2252	0.140	11.727
Midcrustal reflection (oceanic)	133	0.034	0.477
P_mP /top underplate	1066	0.251	8.860
Base underplate	131	0.069	1.937
P_n	427	0.147	3.372
All phases	11991	0.110	3.662

the model only if required by the data and layers are only included if reflected arrivals or changes in the velocity gradients are necessary to explain all arrivals. Velocity gradients and the phase identification in the velocity model were further constrained by synthetic seismogram modeling using asymptotic ray theory [Zelt and Ellis, 1988]. Picking of the onset of first and later arrivals was performed without filtering where possible (mostly between offsets of 0–40 km). Different filters were applied to the data where necessary, depending on the quality of the data and offset to the source.

[18] Estimated picking uncertainties were 40 ms for sedimentary and upper crustal turning ray arrivals, increasing to 60 ms for middle to lower crustal turning rays and reflected arrivals. The largest uncertainties were assigned to phases identified as resulting from diving rays into the upper mantle (P_n) or reflected on the Moho ($P_{m2}P$) or on the top of the igneous underplate ($P_{m1}P$). The number of picks, root-mean-square (RMS) travel time residual, and the

χ^2 error for all phases are listed in Table 1 for line E and in Table 2 for line D.

[19] The final velocity model of line E consists of nine layers: the water layer, two sedimentary layers, one layer of basalt, three crustal layers, an underplate layer, and the upper mantle layer. Each layer is defined by depth and velocity nodes. Water velocity is a constant 1500 m/s throughout the model, which agrees well with existing water velocity data [Dietrich *et al.*, 1975]. Seafloor bathymetry was determined from towed hydrophone data, due to the lack of digital echo sounder logs. The seafloor model layer includes depth nodes at a spacing of 2.5 km (Figure 8). The two sedimentary layers found over oceanic crust and in the basin are modeled with the same node spacing. Sediment velocities range from 1.9 to 3.8 km/s. In the trough a layer of basalt with a thickness up to 1.5 km has been modeled from the maximum distance of turning rays arriving from this layer and its amplitude with a velocity node spacing of 2.5 km overlying a third sedimentary layer

**Figure 8.** Final velocity model of line E including the model boundaries used during inversion (solid lines) and isovelocity contours every 0.25 km/s. Positions of OBSs (inverted triangles) are indicated.

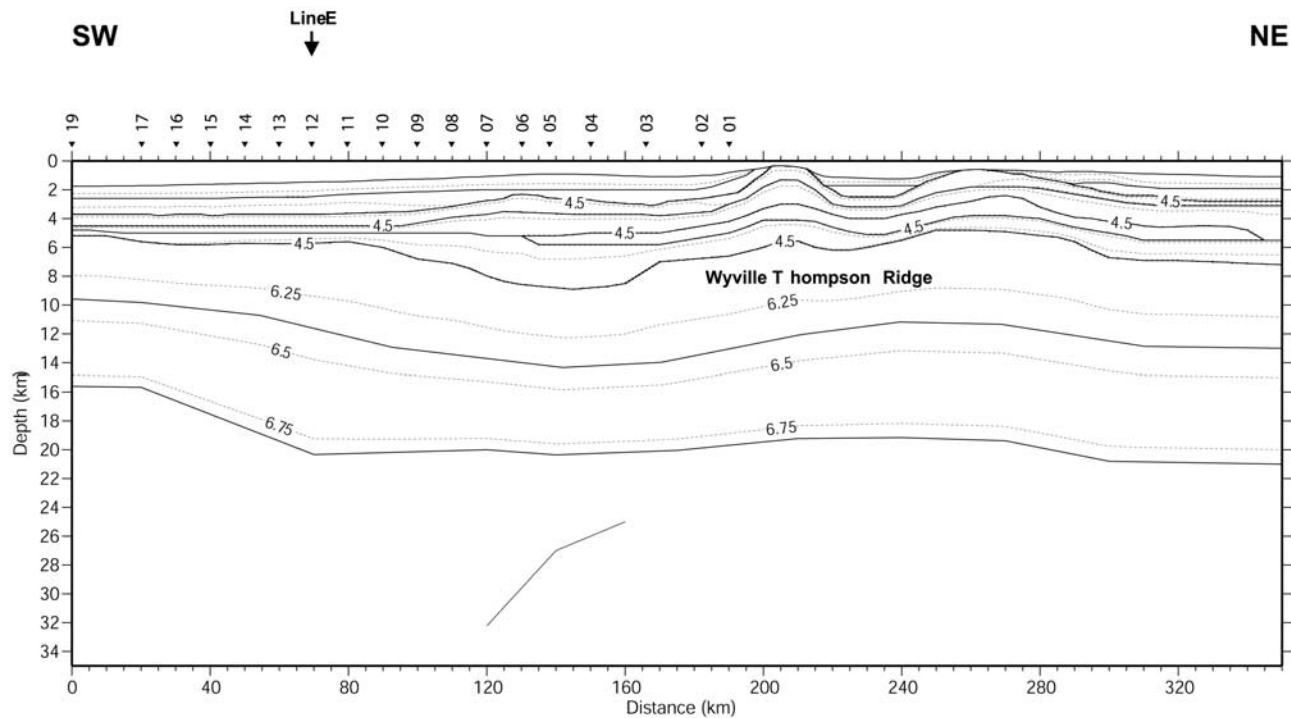


Figure 9. Final velocity model of line D and BP/Shell data sets including the model boundaries used during inversion (solid lines) and isovelocity contours every 0.25 km/s. Positions of OBSs (inverted triangles) are indicated.

with the same node spacing. The velocity of the subbasalt sediments could not be constrained, as this layer forms a low-velocity zone and hence turning rays originating from this layer are not observed. The velocities of 3.00–3.80 km/s were assigned to this layer based on sediment velocities found where basalt was absent. The top of the basement has also been modeled with a node spacing of 2.5 km to account for its roughness. The midcrustal layer, the top of the underplate layer and the Moho are less well resolved in the data and a depth node spacing of 10–12 km was adequate. Velocity nodes are only used when required by the model to account for lateral velocity variations, e.g., at the ocean-continent transition zone where normal oceanic crust with high-velocity gradients changes to continental crust displaying much lower-velocity gradients. A layer with velocities between 7.40 and 7.60 km/s, constrained by reflections from its top and bottom has been included in the model west of Lousy Bank and underneath the outer Hebrides shelf. Arrivals from diving rays into the upper mantle were only observed on the land station data and the upper mantle velocity is inferred to be constant throughout the model.

[20] The final velocity model of line D includes 10 layers, the water layer, two layers of sediments, a basaltic layer, an intermediate layer of sediments on top of a sill-like layer, a fourth layer of older sediments, two crustal layers and the upper mantle layer (Figure 9). The water velocity is identical to that of line E. The upper layer of sediments displays a velocity of 1.6–2.2 km/s, the second layer a velocity of 2.35–2.70 km/s. The boundary between the two layers is constrained by reflected arrivals from the boundary between them. Seafloor bathymetry from the towed hydrophone data

and both sedimentary layers have been modeled using a node spacing of 2.50 km. The basaltic layer is generally thinner than on line E, thinning eastward from around 700 m thickness to only 300 m thickness. The velocities are typical for basalt, around 4.20–4.80 km/s. The sediments underneath the basaltic layer have been modeled with velocities from 3.95 to 4.15 km/s and therefore represent a low-velocity zone. Between model distance 130 km and 340 km a second high-velocity layer with a subsequent velocity inversion to the underlying sedimentary layer has been modeled. The deepest sediments were modeled with a velocity of 4.20 to 4.70 km/s. Node spacing of the basaltic and lower sedimentary layers was 5 km. No reflections from the midcrust interface have been found in the data so that no velocity discontinuity in the crust has been incorporated into the model. The crust consists of two layers, the upper crust displaying a velocity gradient from 6.0 to 6.4 km/s and the lower crust from 6.4 to 6.8 km/s. However, a small discontinuity may have remained undetected due to the data quality. Arrivals reflected from the Moho and from diving waves into the upper mantle have been identified throughout the data set. The upper mantle velocity could thus be constrained to 8.0–8.2 km/s. Node spacing at the Moho was around 20 km. At around 140 km model distance a deep reflection observed on four instruments has been included in the model as a mantle reflector.

4. Error Analyses

[21] Velocity gradients and the phase identification were constrained by synthetic seismogram modeling [Zelt and Ellis, 1988] (Figures 3, 4, 6, and 7). In the synthetic

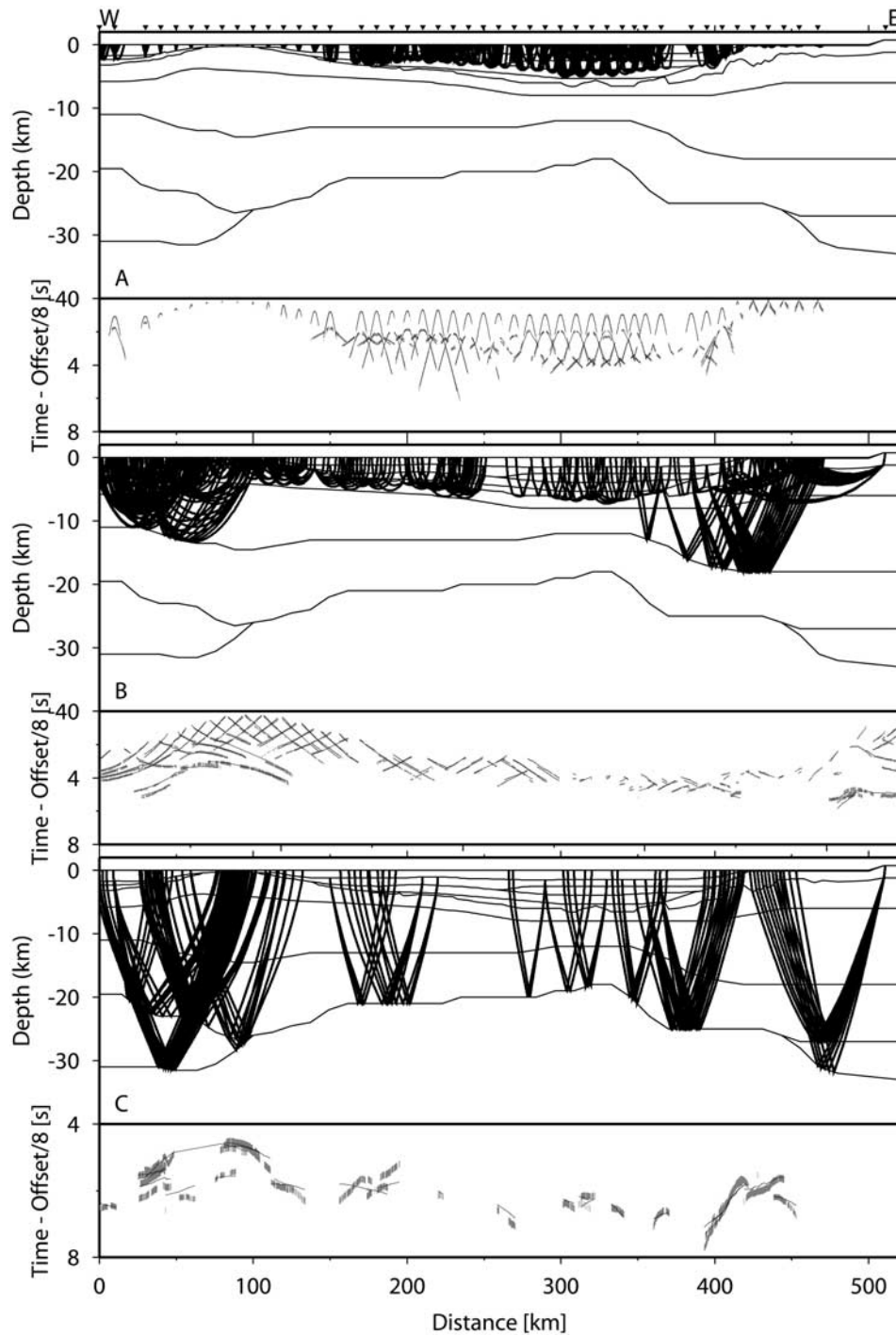


Figure 10. (a) (top) Ray coverage of the model of line E for the sedimentary layers, with every 20th ray from point-to-point ray tracing. Positions of the receivers are indicated by inverted triangles. (bottom) Fit between the travel time picks (dark grey bars) and the predicted arrival times (black lines) from ray tracing for the sedimentary layers. (b) Same as Figure 10a except for the upper crustal layer. (c) Same as Figure 10a except for the lower crustal and upper mantle layers.

seismograms the sedimentary layers typically generate low-amplitude reflections, often overlain by multiple arrivals of the sea bottom reflection. The layer of underplate underneath oceanic crust close to Lousy Bank produces a characteristic double reflection (Figure 3). The basaltic layer, where present, causes a high-amplitude reflection

and the high-velocity gradient in this layer causes a high amplitude arrival out to offsets up to 20 km (Figure 6). The low velocity sediments underlying the basalt cause a characteristic step back in the first arrival times, delaying the crustal turning rays (Figure 6). Since a low-velocity layer does not produce turning rays, the constraints on this layer

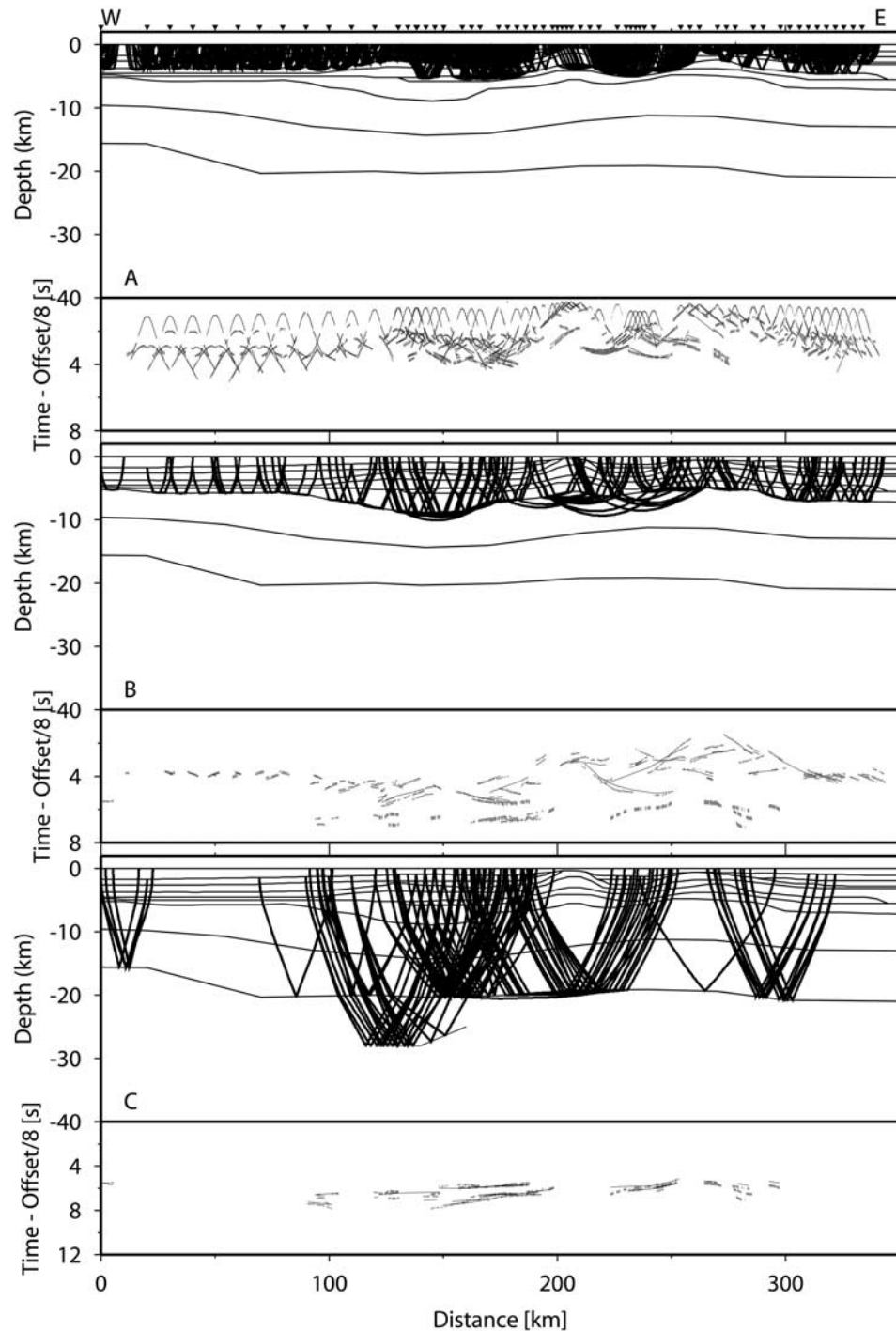


Figure 11. (a) (top) Ray coverage of the model of line D BP/Shell for the sedimentary layers, with every 20th ray from point-to-point ray tracing. Positions of the receivers are indicated by inverted triangles. (bottom) Fit between the travel time picks (dark grey bars) and the predicted arrival times (black lines) from ray tracing for the sedimentary layers. (b) Same as Figure 11a except for the upper crustal layer. (c) Same as Figure 11a except for the lower crustal and upper mantle layers.

are generally weak. The high-velocity gradients of the upper oceanic crust in the model give strong amplitude arrivals (Figure 3), while the lower gradients of the continental crust result in lower-amplitude arrivals (Figure 4). The strong velocity contrasts across the Moho (PmP) lead to a strong reflection visible on most data sections from line E (Figure 4).

[22] On line D a late very high amplitude reflection can be observed, (Figure 6), which has been modeled as a mantle reflector. It is impossible to fit these arrivals as a Moho reflection and simultaneously construct a gravity model which satisfies the observed gravity data. The basaltic layer and the sill-like body give rise to high-amplitude arrivals (Figure 7).

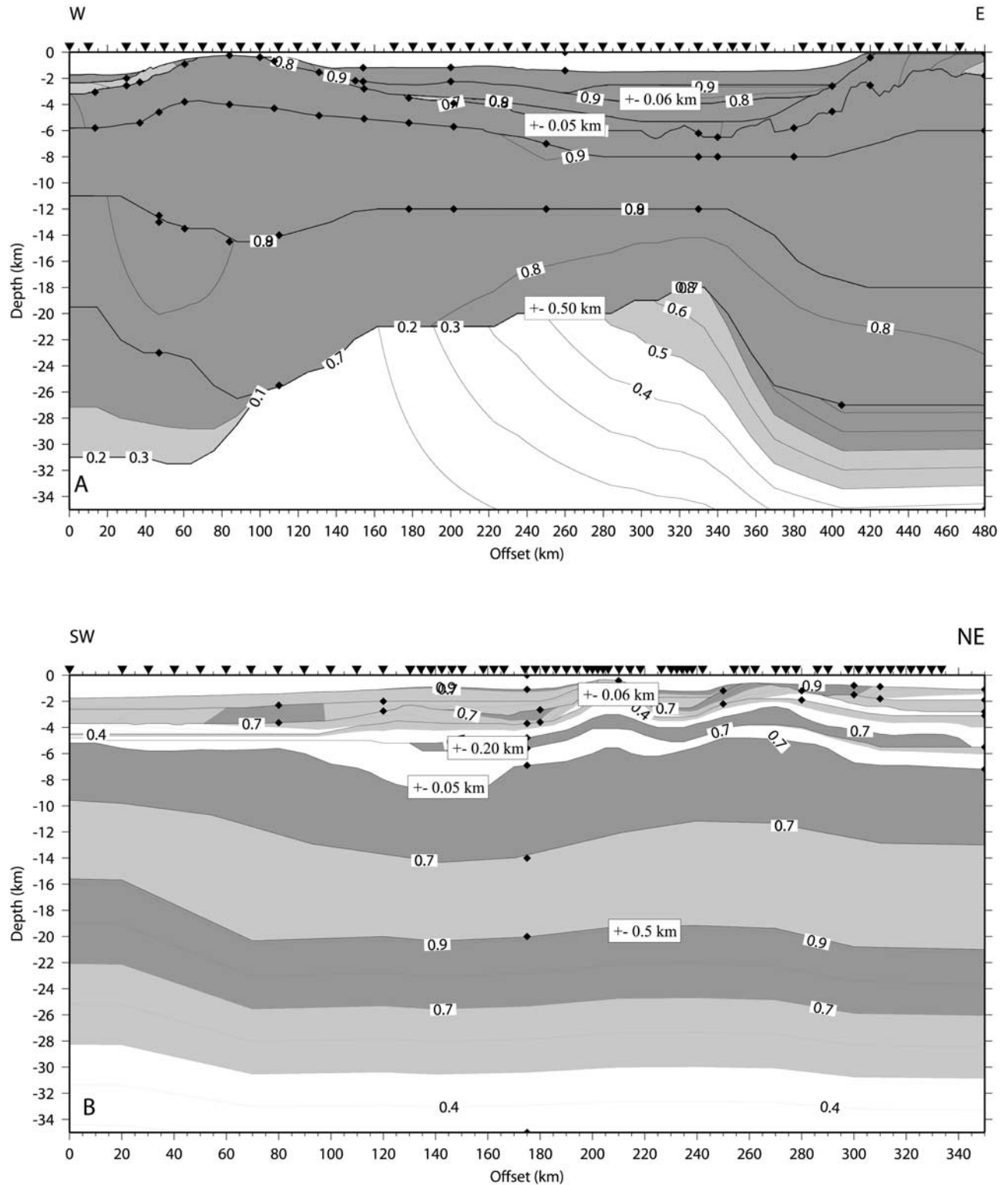


Figure 12. (a) Resolution parameter for all depth nodes of the velocity model of line E. Velocity nodes are marked by black diamonds, and OBS positions are marked by black inverted triangles. Contour interval is 0.1. The depth uncertainty of the most important boundaries calculated from the 95% confidence limit of the F test is given in the framed boxes. (b) Resolution parameter for all depth nodes of the velocity model of line D. Velocity nodes are marked by black diamonds, and OBS positions are marked by black inverted triangles. Contour interval is 0.1. The depth uncertainty of the most important boundaries calculated from the 95% confidence limit of the F test is given in the framed boxes.

[23] Two-point ray tracing between source and receiver (Figures 10a and 10b) shows well resolved and unconstrained areas. On line E the sedimentary layers and upper basement are well resolved throughout the model. Fewer reflected rays are seen from the midcrustal and Moho discontinuity, the only major gap with no ray coverage of the Moho is between model distance 100 and 170 km and smaller gaps exist between 190–230 km and 340–370 km model distance. The depth of a midcrustal velocity discontinuity is constrained by reflections between 0 and 80 km model distance and at distances greater than 300 km. Ray coverage on line D is good between model distance 80 and 350 km, however the Moho is not constrained by seismic data between 10 and 60 km model distance. No reflections from a midcrustal discontinuity could be distinguished in the data, and the boundary represents a second-order discontinuity with no associated velocity step in the model. For both lines the depth of the Moho has been additionally constrained by gravity modeling in those areas of insufficient ray coverage.

[24] The fit between predicted arrival times and travel time picks provides information about the quality of the model (Figures 10a, 10b, and 10c, bottom, and Figures 11a, 11b, and 11c, bottom). The χ^2 error is defined as the root-mean-square (RMS) travel time misfit between observed and calculated arrivals normalized to the picking uncertainty. The number of picks, picking error, the values for the χ^2 parameter, and the RMS misfit for the most important phases of the models are listed in Tables 1 and Table 2.

[25] Additional information about the quality of the velocity model can be gained from the resolution parameter (see Figures 12a and 12b) [Zelt and Smith, 1992]. Resolution is a measure of the number of rays passing through a region of the model constrained by a particular velocity node and is therefore dependent on the node spacing. If a layer can be modeled with one single velocity gradient the resolution parameter will be high even in areas which have lower ray coverage as the area is related to only one velocity node. The advantage of this representation is that it allows an assessment of whether all lateral velocity changes are required by the data. Hence no structures unconstrained by the data have been included in the model [Zelt, 1999]. Nodes with values greater than 0.5 are considered well resolved (Figure 12a). The upper sedimentary layers and the upper crust of line E are well resolved. Velocities of the low-velocity layer underneath the basalt are not constrained by turning rays, thus this layer shows a low resolution. Although the lower crust is less well resolved than the upper crust, values are still well above 0.5 and can thus be considered reliable. Upper mantle velocities are only constrained by the land station data and the resolution drops toward the western end of the profile. On line D upper sedimentary, basaltic, upper crustal and upper mantle velocities have a resolution greater than 0.5 (Figure 12b). Again the low-velocity zones underneath the basalt and the sills are unconstrained by turning rays and therefore show a lower resolution. The lack of turning rays in the lower crust leads to resolution values of 0.5 which can still be considered as sufficiently well resolved.

[26] In order to estimate the velocity and depth uncertainty of the final velocity model a perturbation analysis was performed. The depths of key interfaces were varied and an

F test was applied to determine if a significant change between models could be detected. The 95% confidence limit gives an estimate of the depth uncertainty of the interface (Figures 12a and 12b). The top of the basaltic layer and the top of the basalt are very well constrained with an uncertainty of ± 60 m. The sill-like body in line D is less well constrained with an uncertainty of ± 200 m due to the fact that it is located below a low-velocity zone. Uncertainty of the Moho is around ± 0.5 km for both models.

[27] To determine the robustness of our conclusions concerning the absence of the high-velocity body underneath the trough, we constructed different velocity models including a model with a layer of underplate at 31.5 km depth, and a model with a layer of underplate of a constant thickness of 4 km and subsequently calculated synthetic seismograms for each model (Figure 13). Both alternative models predict a double reflection from the top and base of the underplate with each a lower amplitude than the single high amplitude reflection predicted by the preferred velocity model, which is more consistent with the data observed.

[28] In order to test whether a small region of oceanic crust is present in the middle of the trough, an alternative velocity model was produced. This model included higher, more oceanic type, velocity gradients in the middle of the trough. Comparison of synthetic seismograms from the final velocity model and the alternative velocity model indicates that the higher-velocity gradients in the oceanic lower crust produce a stronger amplitude arrival than found in the data (Figure 14). We conclude from this that although arrival times found in the trough can be explained by oceanic crust as well as continental crust, the amplitudes of crustal arrivals are better explained by continental type crust. However, anomalous (highly intruded) or oceanic crust with an extent smaller than the OBS spacing cannot be excluded on the basis of the velocity modeling.

[29] The different number of crustal layers between the two velocity models (lines E and D) at their intersection is a result of the minimum parameter/minimum structure modeling approach. We use only the smallest number of layers necessary to explain all travel time picks within their error bounds. As the velocity steps required to fully fit the data are relatively small on line E (6.00 to 6.20 km/s and 6.40 to 6.60 km/s) it might not be possible to detect them on line D. Small differences in top basement and Moho depths at the intersection of the lines are within the error bounds of the given interfaces (Figure 12).

5. Gravity Modeling

[30] Assuming seismic velocities and densities can be correlated, gravity modeling provides important additional constraints on the seismic model. Areas unconstrained by the seismic data can be modeled by comparing calculated gravity anomalies with those observed. Crustal P wave velocities from the seismic models were converted to densities using the relationship of Christensen and Mooney [1995] with upper mantle densities set to a constant 3.32 g/cm^3 . Sedimentary velocities were converted to densities using the different empirical relationships for sedimentary layers [Hamilton, 1978; Hughes et al., 1998; Ludwig et al., 1970] (Figure 15). Gravity anomalies along the profiles were extracted from the satellite derived gravity data set of

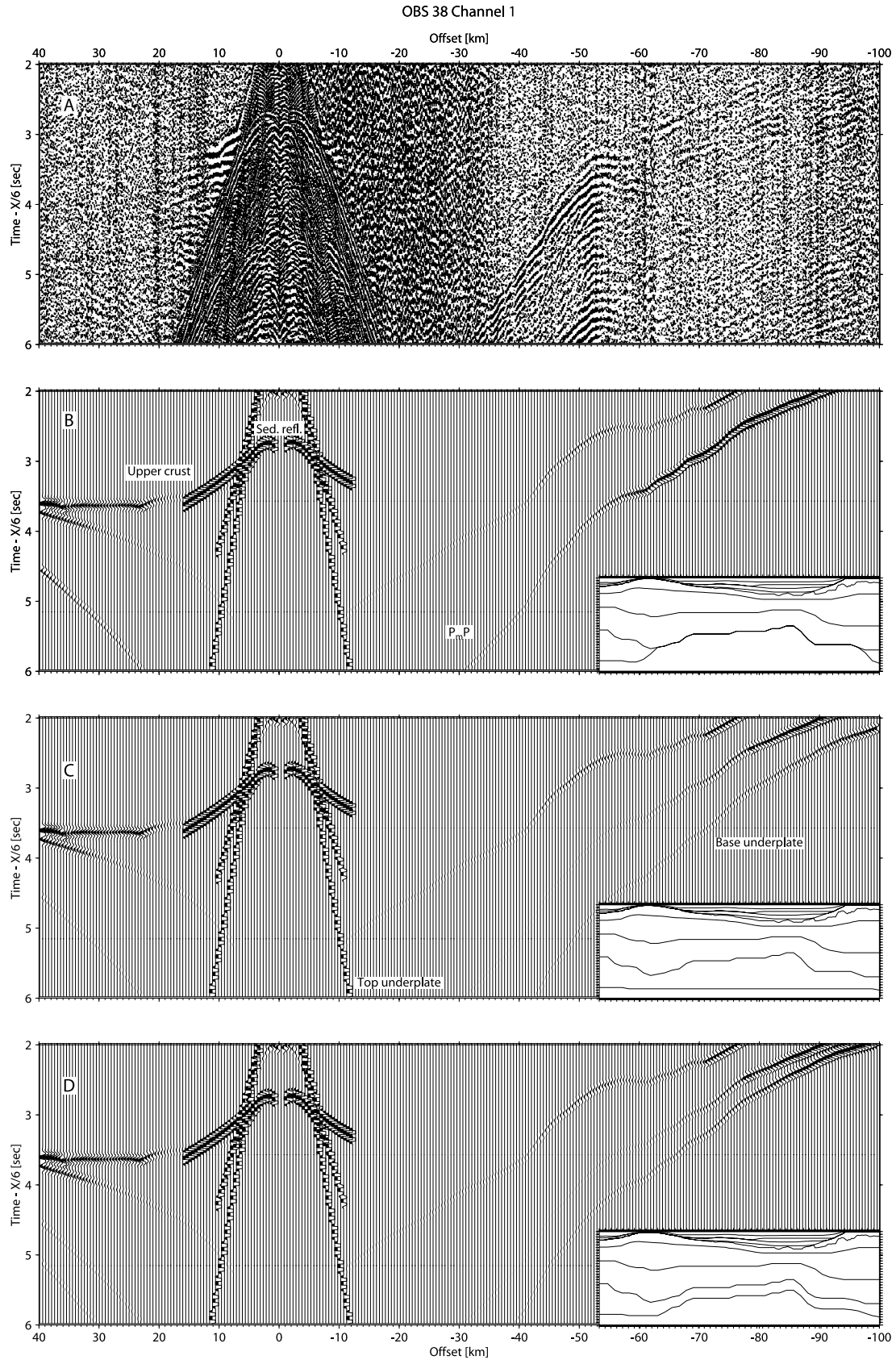


Figure 13. (a) Vertical geophone section of OBS 38. (b) Synthetic seismograms calculated from the final velocity model shown in inset. (c) Synthetic seismograms calculated from a velocity model with a layer of underplate at 31.5 km depth. (d) Synthetic seismograms calculated from a model with a layer of underplate of a constant thickness of 4 km.

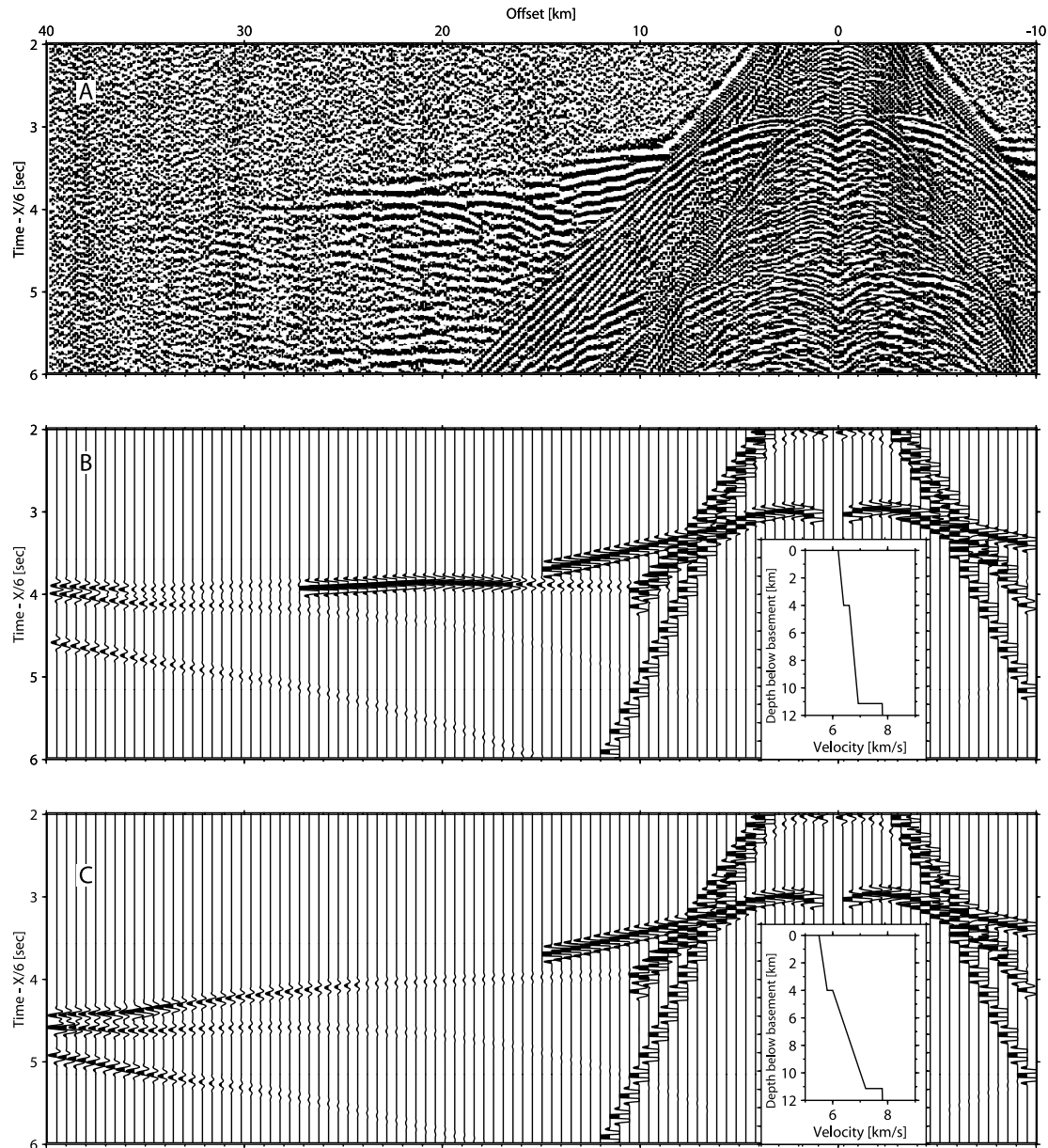


Figure 14. (a) Vertical geophone section of OBS 35. (b) Synthetic seismograms calculated from the final velocity model (velocity depth function at the OBS is shown in inset). (c) Synthetic seismograms calculated from a velocity model with oceanic type velocity gradients (velocity depth function at the OBS is shown in inset).

[Sandwell and Smith, 1995] (Figure 16). The gravity data were forward modeled using the 2.5D GRAVMAG gravity and magnetic modeling software developed by the British Geological Survey [Pedley *et al.*, 1993]. To avoid edge effects both models have been extended by 100 km at both ends and down to a depth of 95 km.

[31] The line E gravity model consists of 17 polygons (Figure 17). Densities used for the modeling are shown in Figure 15. The background density has been set to 2.94 g/cm^3 , the average density of the model. The upper crust of the gravity model has a density of 2.80 g/cm^3 , densities on top of Lousy Bank are slightly less, corresponding to the lower seismic velocities. The density of the lower crust in the oceanic part of the model is 3.05 g/cm^3 . Lower densities

have been assigned to the lower crust in the center of the trough. Underneath the Outer Hebrides Shelf the model requires higher densities. This is reflected in the higher crustal velocities in this region. The base of the crust displays velocities up to 7.2 km/s which translate into densities of 3.05 kg/m^3 [Christensen and Mooney, 1995]. A possible explanation for this slightly elevated velocity and density might be an intrusion in the lower crust or a layer of magmatic underplate, which grades into the lower crust rather than having a sharp boundary, as only few reflections from the top of the possible underplate layer are found in the seismic data in this region, in contrast to the NW end of the line. Here the best fit to the observed gravity can be obtained by assigning a density of 3.1 g/cm^3 to the

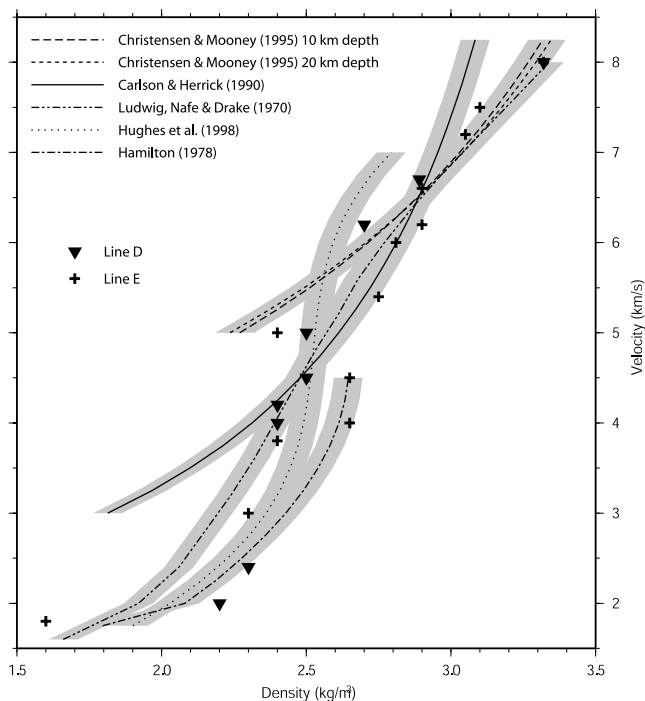


Figure 15. Relationship between velocity and density from various publications. Crosses and inverted triangles mark velocity and corresponding densities used for gravity modeling of both profiles. Grey shaded areas mark the error bounds of 0.05 kg/m^3 .

high-velocity layer found under the oceanic part of the data and Lousy Bank itself. No indications are found from either the seismic or the gravity modeling for a thick layer of underplate beneath the center of the trough. The fit between the gravity anomaly predicted from the modeling and the free-air gravity anomaly calculated from satellite altimetry

is generally good. The largest discrepancy is found between model distance 120 and 180 km and is probably due to a gravity high slightly south of the profile (Figure 16).

[32] Densities on line D are similar to those on line E (Figure 18). Sedimentary densities were chosen between 2.20 and 2.40 g/cm^3 , the basaltic layer and the sill density as 2.50 g/cm^3 . The upper crustal layer with a density of 2.70 g/cm^3 is a mean density between the two upper crustal layers of line E. The lower crust has a density of 2.90 g/cm^3 similar to the lower crust on lines A and E. The upper mantle has been set to 3.32 g/cm^3 . The overall fit is good and there was no need to divide the seismic layers into different density layers. Both ends of the model show an increasing misfit. At the NE end this is due to the influence of the Westray ridge. At the SW end the misfit is probably due to increasing basement depth.

6. Reprocessing of the BANS Data Set

[33] The ESP and sonobuoy data were forward modeled using the same software as for line E [Zelt and Smith, 1992]. In order to be able to interpolate between the sonobuoys and ESPs the main reflectors were digitized from the confidential coincident deep seismic reflection sections, including up to three sedimentary layers, a basaltic layer, basement and in some Moho reflections. The travel times were converted to depth using the seismic velocities from the sonobuoy and ESP data and the reflectors were subsequently incorporated into the wide-angle seismic modeling (Figure 19a).

[34] The fact that sonobuoys are subject to drift with the prevailing ocean currents after deployment can introduce significant travel time errors into the modeling. The true offset between the sonobuoy and the ship can be calculated using the direct arrival traveling through the water column with the precisely known water velocity. It is then possible to recalculate the offsets for the complete section. One

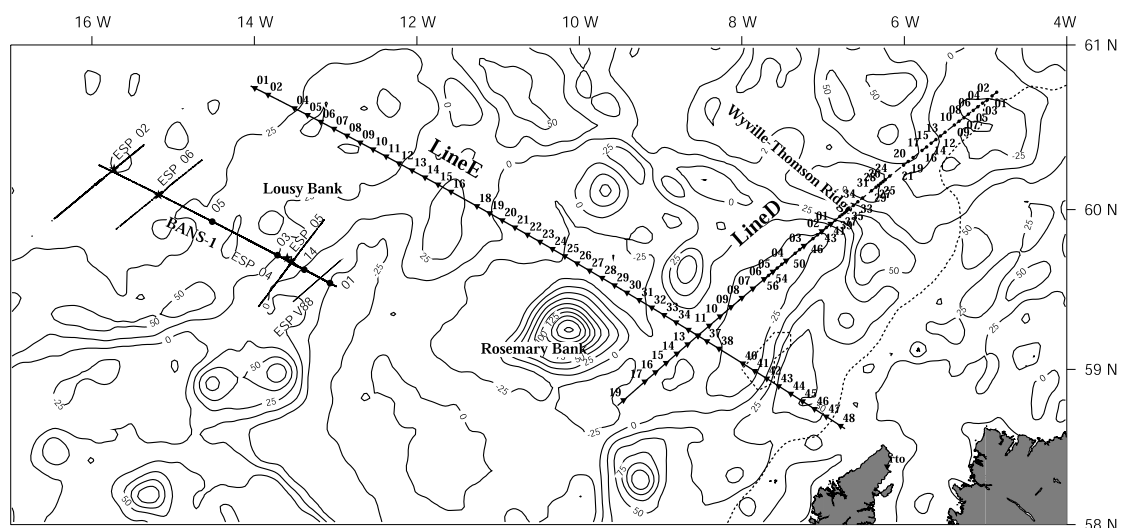


Figure 16. Free-air gravity anomaly in the NE Rockall Trough from satellite altimetry [Sandwell and Smith, 1995], contoured every 25 mGal . The gravity anomaly south over Rosemary Bank south of line E causes a discrepancy between measured gravity by satellite and predicted gravity anomaly from modeling.

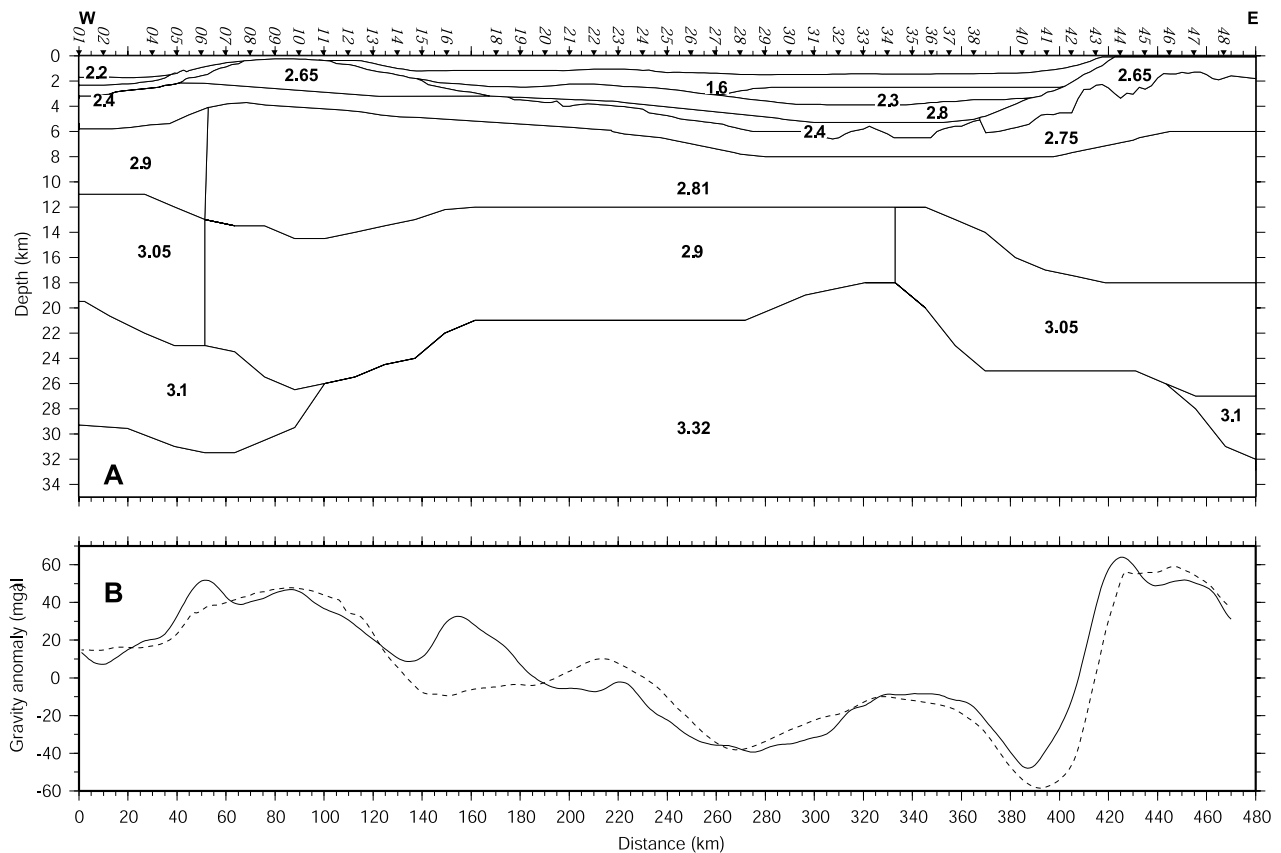


Figure 17. Results of the gravity modeling on line E. (a) Density model. Crustal P wave velocities from the seismic models were converted to densities using the relationship of *Christensen and Mooney* [1995] and others (Figure 15) with upper mantle densities set to a constant 3.32 g/cm^3 . Unconstrained parts of the velocity model have been changed to improve the fit of the gravity. (b) Fit between the predicted gravity anomaly (dashed line) and the measured free-air anomaly from satellite altimetry (solid line) on line E.

assumption made was that drift was linear at a constant speed throughout the active life span of the instrument. Using this simple correction it was possible to reduce the average travel time error for the direct water wave and the reflections from the seafloor from 150 ms to 80 ms, on all profiles. The signal/noise ratio in the sonobuoy section is generally lower than in an OBS section due to the wave generated noise being sourced closer to the instruments and the fact that the sound waves have to cross the water column twice. Data quality on the instruments used for the modeling is generally good for the sonobuoy data, however, on the BANS profile no deep crustal arrivals could be picked.

[35] For the expanding spread profiles, the shooting and receiving vessels start from a common midpoint and travel apart maintaining equivalent shot and receiver distances along the track. As the distance between the ships increases the first arrival sound wave originates from greater depths. Explosives were used as the source with an increasing charge from 10 kg to 400 kg along each 92 km long profile. Usually expanding spread profiles are modeled as one dimensional velocity depth profiles, assuming structure is 1-D underneath the midpoint of the profile. Accordingly, the profiles should be shot parallel to any bathymetric or structural feature, rather than across it, in order to find true seismic velocities. While BANS ESPs 02, 05 06 and V88-6

were shot parallel to the margin, ESP 04 was shot across the margin, where clearly the assumption of one dimensionality is not valid.

[36] In order to be able to allow for varying bathymetry on the ESP 04 for which the one dimensionality assumption is not valid and to include 2-D crustal structure, the shots of each ship were binned into 9 average shot points with 10 km distance between them, gradually increasing from the starting shot point. Real shot-receiver offsets were used for the modeling, with the bin shifting along with the shooting vessel, in order to keep the deepest penetrated point underneath the midpoint of the profile. This method allows varying subsurface structure to be taken into account. ESP 02, 05, 06 and 88 for which the one dimensionality assumption is likely to be valid, were modeled one dimensionally using the same software and the final velocity depth profiles were subsequently incorporated into the main profile at the intersection points.

[37] Data quality for all of the ESP is very good with clear arrivals up to the maximum shot offset of 70 km. ESPs 06 and 02 display a clear double reflection around 6s reduced travel time (Figure 20). Where previously seen this has been interpreted as indicative of the presence of underplate [White *et al.*, 1987b]. Unfortunately a mute has been applied to the BANS ESPs, which has erased the first arrivals at

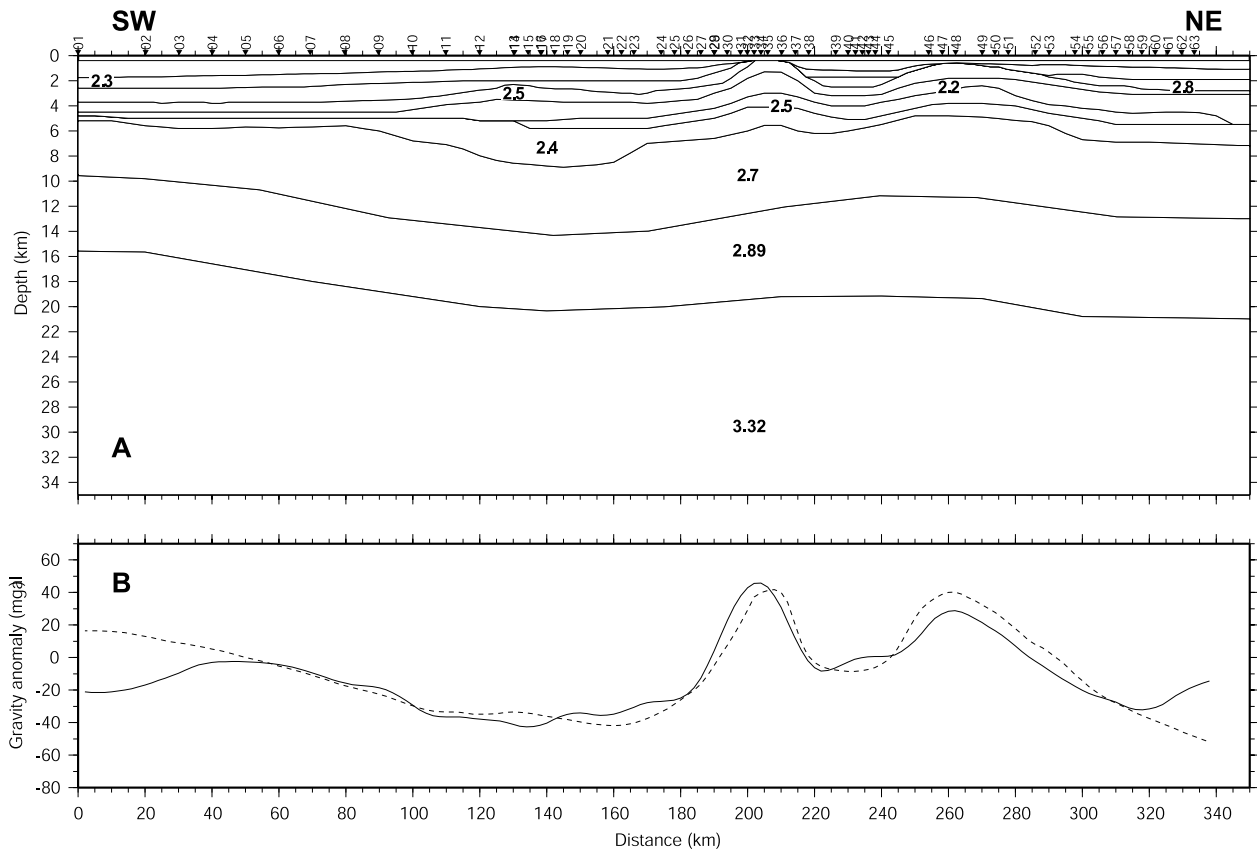


Figure 18. Results of the gravity modeling on line D. (a) Density model. Crustal P wave velocities from the seismic models were converted to densities using the relationship of *Christensen and Mooney* [1995] and others (Figure 15) with upper mantle densities set to a constant 3.32 g/cm^3 . (b) Fit between the predicted gravity anomaly (dashed line) and the measured free-air anomaly from satellite altimetry (solid line) on line D.

offsets between 20 and 30 km, so that the shallow crustal velocities are mainly constrained by the sonobuoy data.

[38] Because of the lower signal/noise ratio of the sonobuoys compared to OBSs or ESPs and to allow for travel time errors due to drift remaining in the data after correction, the picking error for the sonobuoy data was set to 80 ms for all phases, which is slightly higher than for the ocean bottom instruments. The same picking error was used for the ESPs. For the longer offsets the 80 ms error is comparable to that of ocean bottom seismometers. The increase in noise due to the double crossing of the water column is compensated by the stacking of the streamer traces. The number of picks, RMS travel time residual and the χ^2 error for all phases are listed in Table 3 for the BANS 1 and the 1-D ESP profiles.

[39] Because of the greater instrument spacing and poorer data quality of the sonobuoys, only five layers have been distinguished and are shown in the final velocity models for the BANS profile (Figure 19b). These include one sedimentary layer (2–2.5 km/s), a basaltic layer with velocities from 4.5 to 5.5 km/s, and two crustal layers (5.60–6.00 km/s and between 6.00 and 6.40–6.60 km/s), a high velocity body (7.00 to 7.40 km/s), and an upper mantle layer (Figure 19b). The high-velocity body has been modeled from a double reflection, seen in ESP 02 and ESP 06 data sections (Figure 20). Interpolation between the two ESPs was

achieved using gravity modeling. No indications of this body could be detected in ESPs 04, 05, and V88-6, although all of them show a clear reflection from the Moho. Therefore this body was not included in the model from the ESP data. Upper mantle velocities were not constrained by the seismic data and were set to 8.00 km/s.

[40] Although the model is less well constrained than the line E and D velocity models and ray coverage is sparse in some regions (Figure 19c), the existence of a high-velocity body underneath the ocean-continent transition zone is well constrained by the double reflection seen in two ESP. To further resolve the crustal structure and the thickness of the underplate in areas unconstrained by wide-angle seismic data, gravity modeling has been performed using the approach described for lines E and D. The resulting fit between the predicted anomaly and the observed satellite gravity data is good (Figure 19e), resulting from the fact that the seismic constraints were less tight leaving more space for changes to fit the gravity anomaly. The slight residual misfits could be due to 3-D or edge effects.

7. Results and Discussion

[41] At the western end of line E two sedimentary layers of 1.5 km thickness overlie a layer with velocities typical of basaltic rocks (Figure 21). On Lousy Bank, where this layer

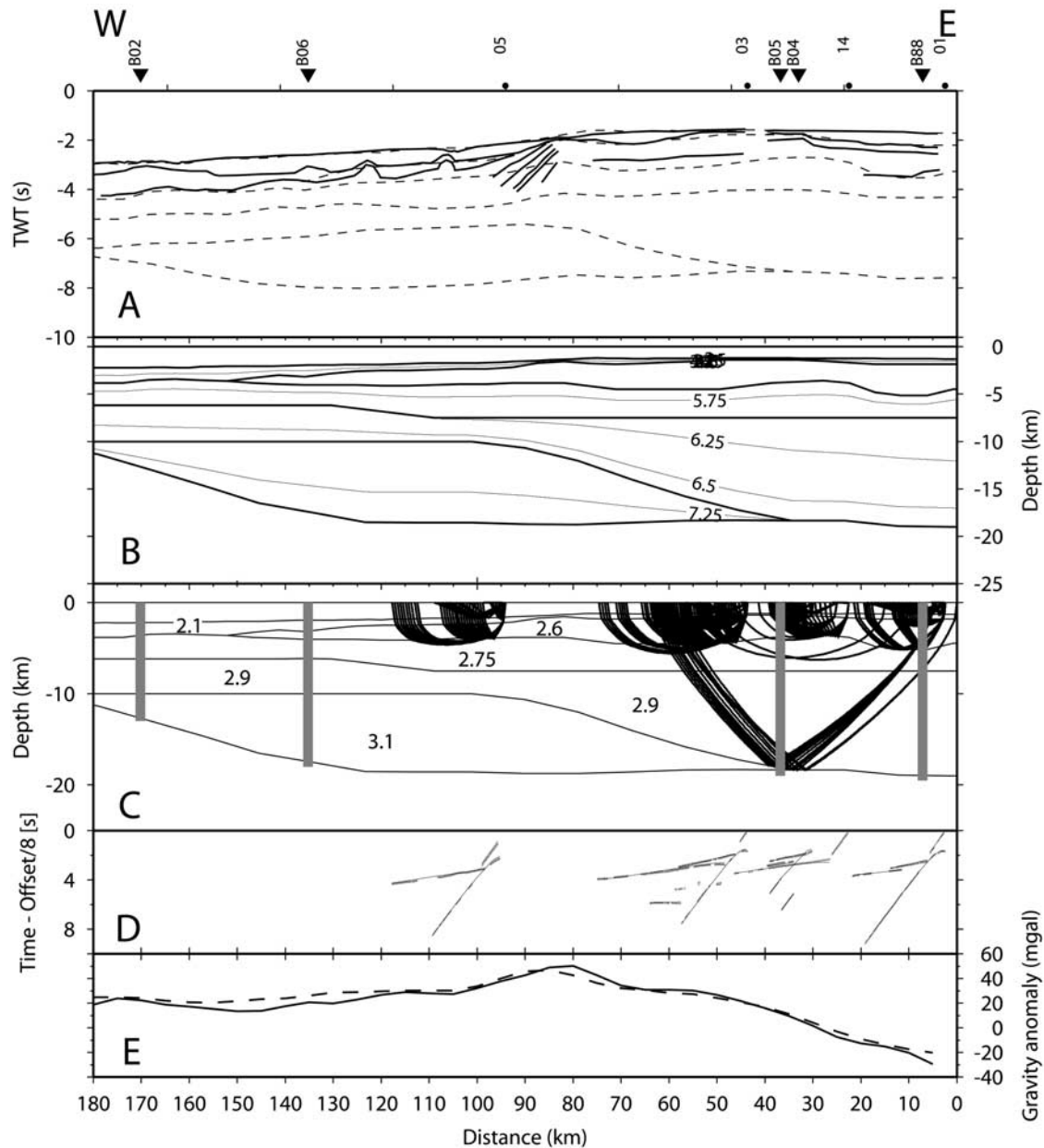


Figure 19. (a) Digitized reflectors from reflection seismic section of the BANS profile (solid lines) and boundaries from the velocity model converted to two-way travel time (dashed lines). (b) Final velocity model including the model boundaries used during inversion (solid lines) and isovelocity contours every 0.25 km/s. Positions of sonobuoys (circles) and ESPs (inverted triangles) are indicated. (c) Ray coverage of the model of the BANS profile for the sedimentary layers, with every 20th ray from point-to-point ray tracing. Vertical bars indicate the cross points with the perpendicular ESPs. Density values used in the gravity modeling are annotated. (d) Fit between the travel time picks (dark grey bars) and the predicted arrival times (black lines) from ray tracing for the sedimentary layers. (e) Fit between the predicted gravity anomaly (dashed line) and the measured free-air anomaly from satellite altimetry (solid line). Density values used in the gravity modeling are annotated in Figure 19c.

crops out analog echo sounder recordings show a rough seafloor typical of basaltic lava flows, which can be correlated with the top of the basalts to the west. Beneath the sediments, a two layer igneous crust is imaged, with velocity gradients and relative thicknesses characteristic of oceanic crust. However, its total thickness of ~ 15 km is unusually thick for oceanic crust [White *et al.*, 1992]. This

might be explained by thick crust formed at the beginning of seafloor spreading in the presence of a mantle plume [White and McKenzie, 1989]. Oceanic crust along the Hatton margin to the SW exhibits a comparable thickness of 11 km [Morgan and Barton, 1990]. Underneath the oceanic crust, an 8 km thick layer displaying velocities higher than normal crustal velocities but lower than normal

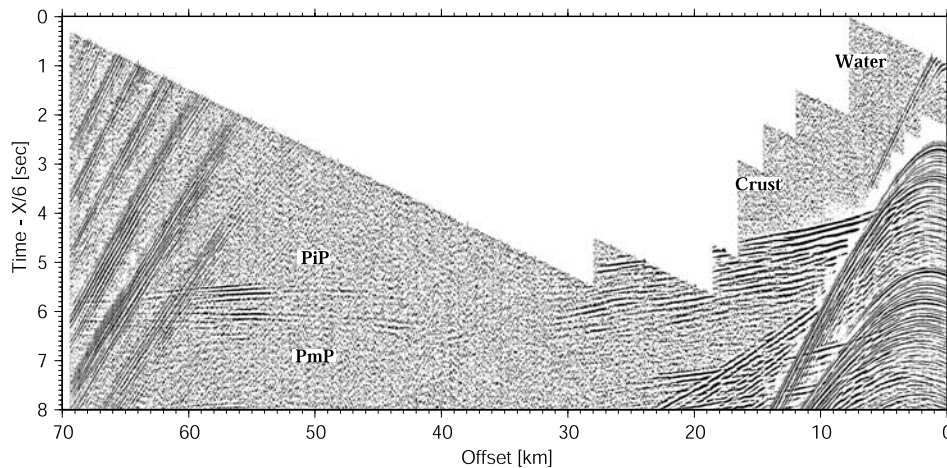


Figure 20. Data from BANS EPS 06 at ocean-continent boundary, showing a double reflection characteristic of an igneous underplate body.

mantle velocities has been imaged. This layer is interpreted to be igneous underplating from the initial rifting. Bodies of similar size and seismic velocity have been described below the Hatton Bank margin [Morgan and Barton, 1990]. These velocities correspond neither to mantle material nor to normal oceanic gabbro which has velocities of 6.8–7.2 km/s. Underplated gabbros derived from melting of asthenosphere with unusually high potential temperature can have increased velocities [White and McKenzie, 1989]. Alternatively, the high-velocity material might correspond to gabbro sills alternating with dunites [Canales *et al.*, 2000] or to olivine-rich gabbros resulting from fractionation of basaltic melt [Farnetani *et al.*, 1996].

[42] The character of the crust changes toward Lousy Bank, where the velocity gradients are shallower and the crust is divided into two layers of roughly similar thickness with velocities between 6.2–6.4 km/s at the Hebrides Shelf end of the profile and 5.75–6.40 km/s at Lousy Bank for the upper crustal layer and 6.6–6.8 km/s at both ends of the line for the lower crustal layer. These velocities are more characteristic of continental crust than of a volcanic edifice. The lower velocities near the top of the crust at Lousy Bank can be explained by lower burial pressures allowing joints to remain open. In addition to the similar velocities found beneath the Outer Hebrides Shelf and Lousy Bank the maximum crustal thickness at Lousy Bank is 26 km, only slightly thinner than underneath the shelf. From this it is

concluded that the crust at Lousy Bank is of continental origin, which is also in good agreement with previous work which found Rockall Bank south of Lousy Bank to be continental in nature [Roberts *et al.*, 1988]. In contrast, oceanic plateaus usually display units having velocities typical of oceanic layers 2 and 3 with velocities corresponding to layer 3 occurring at shallower depth than in continental fragments [Carlson *et al.*, 1980].

[43] In the center of the trough, the sedimentary layers thicken from 2 to 5 km. A thick layer of basalt (up to 1.5 km) can be found in the trough at depths of 2–4 km. The basalt is underlain by a sediment layer up to 2 km thick. The crust thins asymmetrically to around 12 km thickness toward the center of the trough, with the shallowest Moho close to the continental Outer Hebrides Shelf area to the east. The nature of the crust itself does not change, displaying gradients typical for continental crust throughout the trough. The midcrustal reflection reaches its shallowest point at the same location, indicating probably that stretching was highest in this area. A very similar asymmetric crustal geometry has been found in the Rockall Trough slightly further to the south [Roberts *et al.*, 1988]. The crustal thickness increases to 32 km at the eastern end of the profile underneath the continental shelf area. Here it displays three layers with a velocity of 6.0–6.2, 6.2–6.4, and 6.6–7.2 km/s, respectively. The thickness and partitioning is comparable with that previously found west of the Outer Hebrides [Powell

Table 3. Travel Time Residuals and Chi-Square Error for All Phases and the Complete Models

	Phase	Number of Picks	RMS Travel Time Residual	Chi-Square
BANS 1	Water	154	0.134	2.805
	Water reflectance	534	0.074	0.853
	Basalt	583	0.065	0.670
	Crust	952	0.081	1.025
	PmP	102	0.053	0.448
	All phases	2325	0.079	0.987
BANS ESP 02	All phases	695	0.093	1.364
BANS ESP 06	All phases	888	0.124	2.403
V 89-6	All phases	419	0.071	0.780

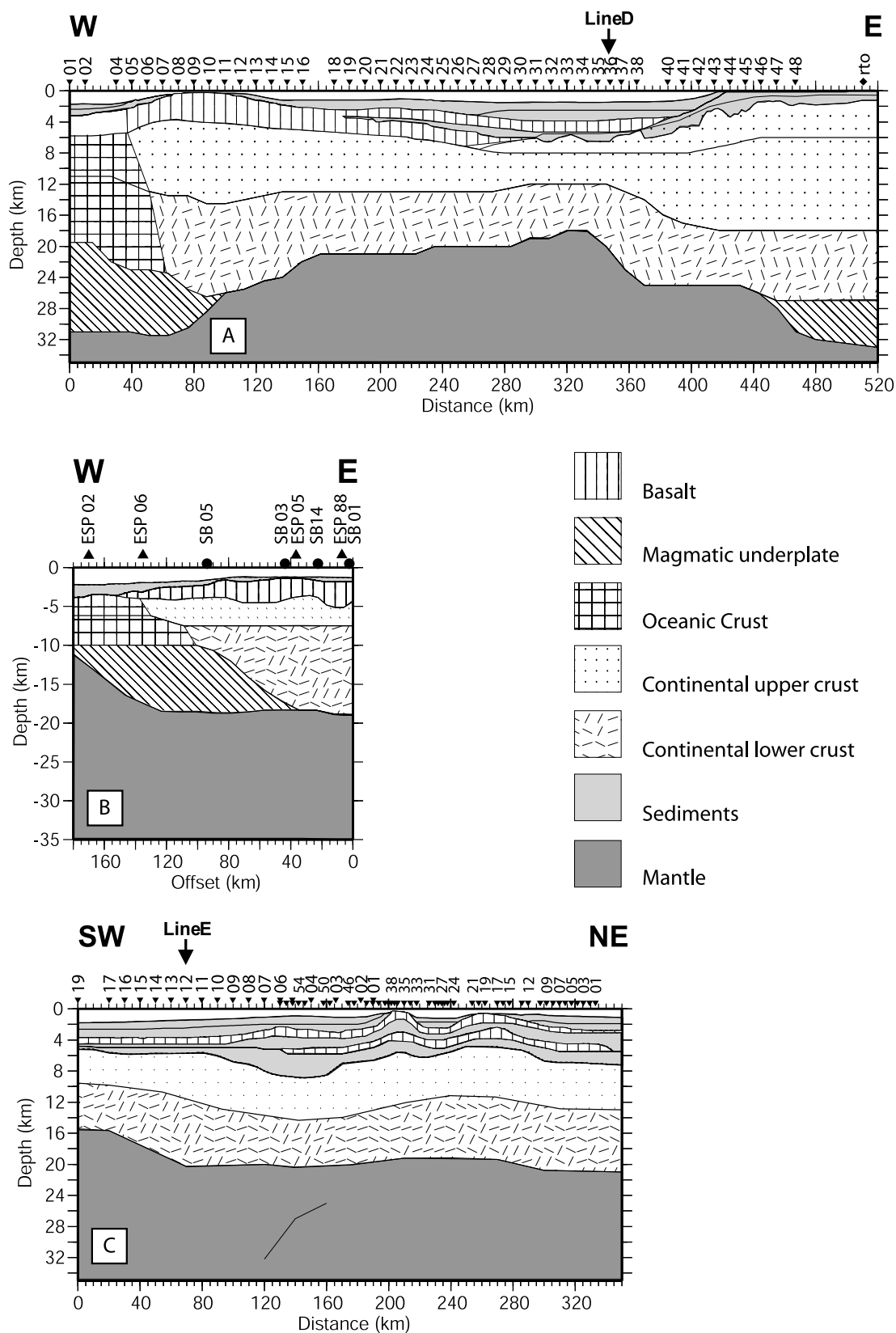


Figure 21. Geological cross sections based on the velocity models of (a) line E (b) BANS, and (c) line D.

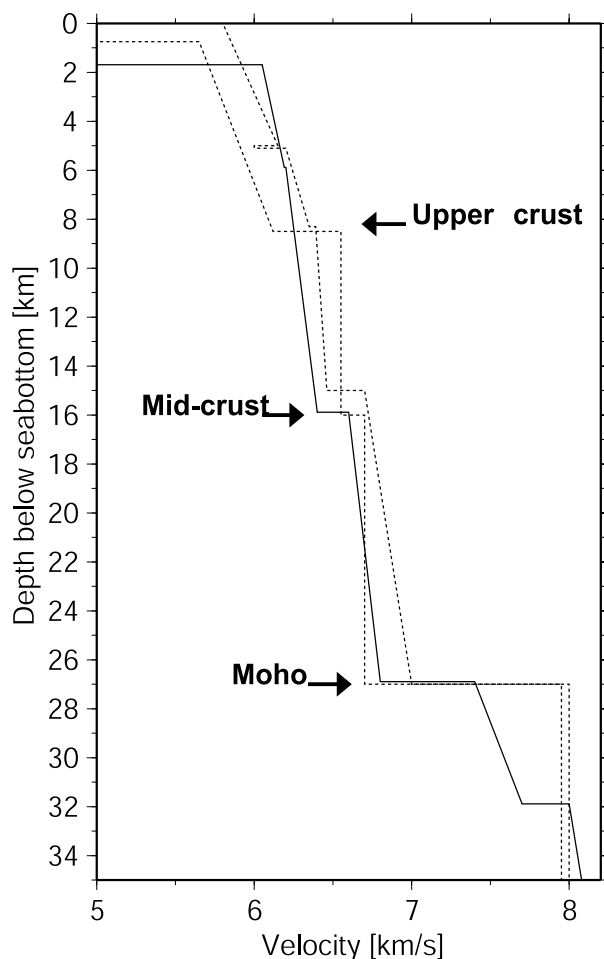


Figure 22. Comparison of velocity-depth profiles for the PUMA experiment (dashed lines) [Powell and Sinha, 1987] and line E (solid line) at 480 km model distance.

and Sinha, 1987] and is believed to be typical of Lewisian basement. The slightly elevated velocity at the base of the crust could be explained by magmatic intrusion into the crust during rifting. A second region of underplating has been modeled underneath the continental shelf, mainly constrained by a double reflection seen on the data recorded by the land station RTO (Figure 5).

[44] The thickness of the sedimentary layer in the BANS profile varies from a few hundreds of meters in the east to two kilometers in the western part of the profile (Figure 21). Underneath the sedimentary layer, a basaltic layer has been modeled, which thins toward the west from 3 km thickness and ends at 100 km model distance. A seaward dipping reflector (SDR) sequence, resulting from basaltic lava flows being erupted on the subsiding continental margin during initial opening of the ocean is imaged in the coincident reflection seismic data at 100 km model distance (Figure 19). The velocity model shows a 18 km thick continental crust at the eastern end of the profile (Figure 21). This crust thins gradually to about 5 km thickness underneath a thick sequence of SDRs at about 100 km model distance. The crust westward of these SDRs is interpreted to be of

oceanic origin. The velocity gradient in the lower crustal layer increases oceanward, typical for the transition to oceanic crust, although the lower crustal velocities might be slightly lower than for typical oceanic crust. This is in good agreement with the normal oceanic crustal thickness of 7 km found in this region. Between 30 km and 180 km model distance the crust is underlain by an up to 10 km thick high-velocity body interpreted to be magmatic underplating similar to the underplate body found on line E.

[45] Sedimentary layers on line D display a thickness of 5.5 km at the SW end of the profile and thicken in the NE Rockall basin to 8 km thickness at 145 km model distance (Figure 21). The thickness decreases toward the NE to about 5 km at 300 km model distance. The thickness of the basaltic layer is less than on line E, and it thins toward the northeast. A second high-velocity layer in the sediments may consist of several sill complexes which cannot be resolved individually. Where the profile crosses the Wyville-Thompson Ridge both high-velocity layers are parallel folded, confirming a compressional origin for the Ridge. Additionally, small basement highs are located underneath the Wyville-Thompson and the Ymir ridges, and no large igneous core (high velocities) is needed to model the ridges, casting doubt on an igneous origin for the Ridges themselves. The crustal thickness along line D increases from 10 to 14 km toward the NE. The velocities and velocity gradients in the crust throughout the model are characteristic of continental type crust. No indications of a high velocity body which could be interpreted as igneous underplating have been found in the model of line D and diving rays into the mantle show, where present, a normal mantle velocity of 8 km/s. A deep high-amplitude reflection is observed between 120 and 160 km model offset. It corresponds in depth to a deep reflection found in the southern Rockall Trough interpreted to be the boundary between serpentinized and normal upper mantle material. However, mantle velocities above the reflector are around 8 km/s on line D. This excludes igneous underplating as well as serpentinization of the upper mantle as possible explanations for the origin of this reflector. This reflector is not imaged on line E, which excludes its lateral continuity between both lines. Because no velocity contrast across this boundary could be observed it is interpreted as an mantle reflection rather than the base of an underplate layer.

8. Comparison to Published Data Sets

[46] The PUMA experiment [Powell and Sinha, 1987] was located off the coast of northern Britain to the west of the Outer Hebrides (Figure 2) over high-velocity, seismically homogeneous, nonreflective, Lewisian basement. A new type of sea bottom receiver, the PUMA, consisting of an 1100 m long array of hydrophones deployed at the seabed was used as the main receiver. The endpoint of the PUMA profile is located about 40 km SW of line E OBS 47. The modeled 1-D velocity-depth relationship from the PUMA data agrees well with the velocity-depth function at the western end of line E (Figure 22). The difference in depth of the midcrustal and Moho reflectors is about 1 km and that of the Moho depth is even smaller. No evidence of underplate was detected in the PUMA data, and an upper mantle velocity of 8 km/s has been modeled using data

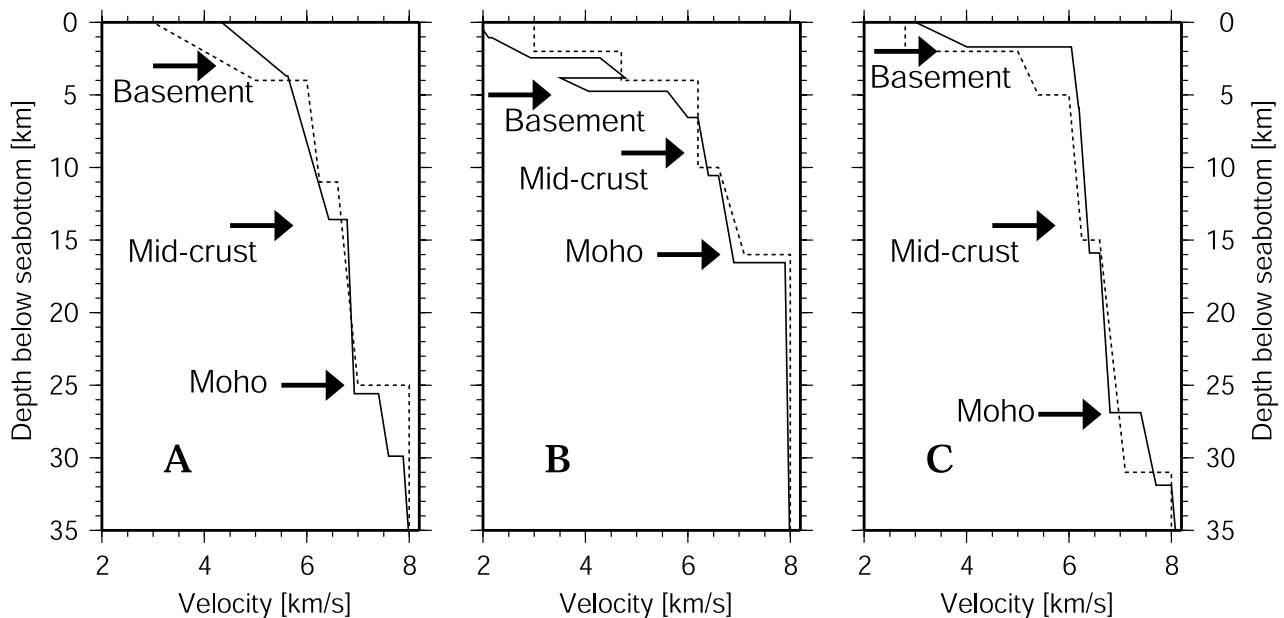


Figure 23. Comparison of velocity-depth profiles for the line P88-2 [Roberts *et al.*, 1988] and line E. (a) Comparison of the velocity depth profile of the line P88-2 (dashed line) and line E at 80 km model distance (solid line) at Lousy Bank. (b) Comparison of the velocity depth profile of the line P88-2 (dashed line) and line E (solid line) at 330 km model distance in the center of the Rockall Trough. (c) Comparison of the velocity depth profile of line P88-2 (dashed line) and line E (solid line) at 480 km model distance at the continental shelf.

collected during the PUMA experiment from two land stations on Lewis.

[47] Roberts *et al.* [1988] present models for two wide-angle seismic profiles across the northern Rockall Trough (Figure 2). The main aims of the project were the determination of the crustal structure and the characterization of the nature of the crust beneath the trough. Profile P2 crosses the Rockall Trough south of line E and the observed crustal structures are similar on both lines (Figure 23). On line P2 an eastward dip of the sediment covered basement and a thinning of the crust on the western flank of the trough is observed, comparable to the thinning of the crust west of Lousy Bank on line E. A general asymmetry of the trough with the shallowest Moho adjacent to the eastern margin of the trough, dipping gently to the west can be found in both models. Crustal thicknesses along both profiles are nearly identical; starting on the western flank with a crustal thickness of about 22 km, thinning in the trough to 11 km and then thickening to around 30 km (Figure 23). No underplating has been detected along profile P2, although areas where underplating could be expected beneath the thicker crust at the margins of the trough are generally unconstrained by the seismic data. Both models show a two-layered crust displaying velocity gradients typical for thinned continental crust and the depth of the midcrustal reflector is comparable in both lines.

[48] Drill holes can provide a correlation between seismic velocities and the lithology of the observed layers (Figure 2). The BP drill hole 154/3-1 is located 70 m from OBS 43 on line E. Drilling penetrated a basaltic layer and ended in Lewisian gneiss. Both the depth of the basaltic layer and depth of the top Lewisian Gneiss agree well with the

velocity model at OBS 43 (Figure 24). A 400 m thick layer of sediments, representing a low-velocity layer underneath the basalt, could not be resolved in the seismic model.

9. Conclusions

[49] Modeling of the wide-angle data along two profiles in the northern Rockall Trough reveals a two-layered thinned continental crust underlying the trough. Velocity gradients between 5.6–6.4 km/s and 6.6–6.8 km/s confirm this. The crustal thickness increases underneath Lousy Bank to about 26 km, only 4–5 km less than undeformed continental crust in this region. From the velocity gradients this is interpreted to be of continental origin rather than oceanic origin and is consistent with knowledge of the nature of the adjacent Rockall Bank [Bullard *et al.*, 1965; Bunch, 1979]. Thin crust with velocity gradients characteristic of oceanic crust have been found west of Lousy Bank suggesting that the ocean-continent boundary lies to the west.

[50] Underneath the ocean-continent boundary a body with velocities between 7.4 and 7.6 km/s has been imaged on profile E and the BANS ESP data. It is constrained in both data sets by a characteristic double reflection from its top and base. This body is interpreted to consist of magmatic underplate, produced by excess mantle melting that accompanied the opening of the NE Atlantic Ocean. Similar bodies have been found further south on this margin [White *et al.*, 1987a; Barton and White, 1997]. However, no evidence for magmatic underplating beneath the NE Rockall Trough has been found in this study, contrary to interpretations based on subsidence analysis [Brodie and

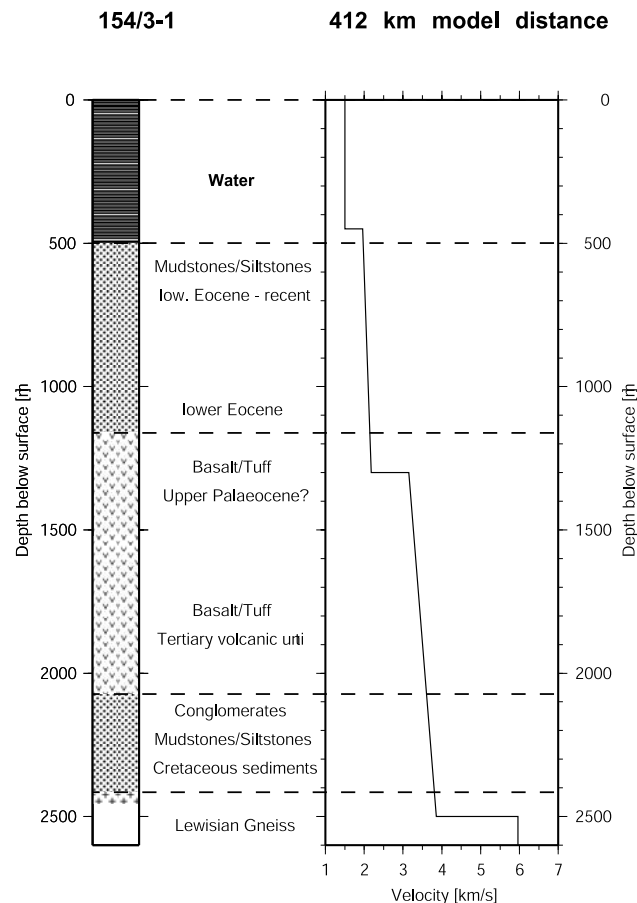


Figure 24. Simplified lithology of well BP 154/3-1 and comparison with the wide-angle velocity model.

White, 1994; Clift and Turner, 1999]. Equally, no evidence for a high degree of serpentinization in the mantle underneath the Rockall Trough has been found as proposed to exist in the southern Rockall Trough [O'Reilly *et al.*, 1996].

[51] Finally, modeling of the structure of the Wyville-Thomson Ridge revealed no existing igneous core of the ridge confirming interpretations of gravity and reflection seismic data [Tate *et al.*, 1999; Waddams and Cordingley, 1999] that this is a compressional structure.

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